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CONTENTS

PAGES

Stratigraphic measurements in parallel folds. By J. B. Mertie, Jr.... 1107–1134

Rock series in diabase sills at Duluth, Minnesota. By George M. Schwartz and Adolph E. Sandberg..... 1135–1172

Glaciation in the Wasatch Plateau, Utah. By Edmund M. Spieker and Marland P. Billings..... 1173–1198

Quaternary history of Louisiana. By Richard Joel Russell..... 1199–1234

Ep-Archean and Ep-Algonkian erosion surfaces, Grand Canyon, Arizona. Robert P. Sharp..... 1235–1270

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

ROCK GLACIERS IN THE COLORADO FRONT RANGE. By Ronald L. Ives.

ORIGIN OF THE MUCK-SILT DEPOSITS AT FAIRBANKS, ALASKA. By Ralph Tuck.

TYPE LOCALITY OF THE CRETACEOUS CHICO FORMATION. By J. A. Taff, G. D. Hanna, and C. M. Cross.

STUDIES OF THE LAKE SUPERIOR PRE-CAMBRIAN BY ACCESSORY-MINERAL METHODS. By S. A. Tyler, R. W. Marsden, F. F. Grout, and G. A. Thiel.

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STRATIGRAPHIC MEASUREMENTS IN PARALLEL FOLDS

BY J. B. MERTIE, JR.

CONTENTS

	Page
Abstract	1107
Introduction	1108
General considerations	1111
Types of folds	1112
Principle of evolutes and involutes	1113
Application of evolutes and involutes	1115
Stratigraphic measurements	1122
General considerations	1122
Several observations along a line	1123
Observations in pairs	1125
Comparison of methods	1132
Works to which reference is made	1133

ILLUSTRATIONS

Figure	Page
1. Sketch illustrating derivation of equation of evolute and involutes, when $\delta = ka$	1115
2. Sketch illustrating measurement of stratigraphic thickness by the use of an evolute and its involutes	1119
3. Diagram showing the type of structure that must exist if $\delta = ka$	1120
4. Diagram showing the derivation of mean trigonometric functions by graphic integration	1123
5. Sketch showing the fallacy of spherical curvature when applied to random observations of strike and dip	1126
6. Sketch illustrating the derivation of $\delta = arc\ tan\ (ks + g)$	1127

ABSTRACT

Folded rocks having bedding surfaces which are approximately parallel are said to lie in parallel folds. Utilizing the principle of evolutes and involutes, the author offers a more precise definition of parallel folds and points out inconsistencies in other concepts. With the idea of classifying parallel folds and possibly of deducing the mechanics of their formation, methods are presented for obtaining the differential equations of the families of involutes which, in certain cross sections, represent the traces of stratigraphic surfaces; and for obtaining the equations of the corresponding evolutes. Geometric methods are also given. These equations and geometric constructions may be deduced either from assumed structural postulates or from actual field data.

Another part of the paper deals with the application of mean trigonometric functions to the measurement of thickness of strata, depth and distance to a stratum, and other stratigraphic dimensions in sections oblique to the strike of the rocks. This topic is considered under two headings: (1) where such measurements can be made from

data collected at several stations along a line of traverse, and (2) where they must be made from a series of structural observations, considered in pairs. In the first case, no assumption is made regarding the curvature of the strata, but instead the mean values of the required functions are derived by mechanical integration. In the second case, the usual assumption of circular curvature is made, and the necessary functions are obtained by the use of definite integrals. Tables of the logarithms of these mean functions, with an increment of 5 degrees for the argument, are also presented.

INTRODUCTION

The measurement of the thickness of strata, the depth and distance to a stratum, and other stratigraphic dimensions constitute a topic that has engaged the attention of structural geologists for many years. Numerous papers have been written on this subject, giving graphic and algebraic methods for the solution of these and related problems. It is not the purpose of this paper to summarize or even to list all these contributions; but a few of them, which have a bearing upon the matters discussed herein, will subsequently be cited. The real object of this paper is to describe new methods of more general scope and application for obtaining stratigraphic information.

Consider, for example, the measurement of the stratigraphic thickness of a series of bedded rocks; and for purposes of brevity, let the significant structural data be defined as follows:

s = the distance between any two designated points in a stratigraphic sequence.
h = the orthographic projection of s onto the plane of the horizon.
e = the difference in elevation between any two designated points, such that $e = (s^2 - h^2)^{\frac{1}{2}}$
σ = the vertical angle, which a line connecting the two designated points, makes with the plane of the horizon.
δ = the angle of dip of a bedding plane.
α = the horizontal angle between h and the angle of strike of a bedding plane.
t = the stratigraphic thickness.

The earliest consideration of this problem was restricted to computations of stratigraphic thickness in a homoclinal sequence of bedded rocks, all of which dipped uniformly in the same direction, within some stated linear interval, say between any two designated stations. Probably many structural geologists developed formulae for the solution of this problem, and some of these formulae were extensively utilized by individual workers long before much had been published on this subject. Professor Charles K. Swartz, of the Johns Hopkins University, for example, drilled his students in such measurements 30 years ago, using a formula of the general type of $t = f(h, e, \delta, \alpha)$. Subsequently various workers pub-

lished papers describing their methods, and among these was one by the writer (Mertie, 1922, p. 41) wherein a formula of the type of $t=f(s, \sigma, \delta, \alpha)$ was derived, together with a nomogram for its graphic solution. The actual formula is

$$t = s\ (sin\ \alpha\ sin\ \delta\ cos\ \sigma \pm cos\ \delta\ sin\ \sigma) \ldots\ldots\ldots\ldots (1)$$

Another formula presented by the writer at the same time, and likewise accompanied by a nomogram, gives the depth and distance to a stratum. This is

$$d = s\ (sin\ \alpha\ tan\ \delta\ cos\ \sigma \pm sin\ \sigma) \cdot \frac{1}{cos\ \rho\ (sin\ \beta\ tan\ \delta \pm tan\ \rho)} \ldots\ldots\ldots (2)$$

where

d = distance to a stratum.

ρ = the vertical angle which a sloping drill hole or tunnel makes with the plane of the horizon.

β = the horizontal angle which the orthographic projection of the sloping drill hole or tunnel makes with the angle of strike of the rocks.

s, σ, δ and α are defined as before.

If the drill hole or shaft is vertical, the fractional member of formula (2) vanishes. These formulae are given herewith because they form the basis for certain topics to be discussed later.

Formulas (1) and (2) presume that the strike and dip of a series of rocks remain constant between any two designated stations and therefore do not apply strictly to folded rocks, where the strike and dip vary from station to station. Nevertheless, because the strike and dip of rocks in many sections change slightly from point to point, and also because many sections are measured in small intervals, these and related formulae have been extensively utilized in the measurement of folded rocks. In practice, the initial and terminal dips of a section, δ_1 and δ_2, and the corresponding values of α_1 and α_2 are averaged to give a mean δ and a mean α, which are then applied in formulae (1) and (2). This method assumes that δ and α vary uniformly with the distance s, or, in other words, that $\delta = k_1 s$, and $\alpha = k_2 s$, when k_1 and k_2 are easily determined constants. This is one of the many assumptions that might be made regarding the relations between δ and s and between α and s, and, as later shown, this assumption may in some sections be entirely justified. Some conclusions regarding the character of the folding under this assumption are also presented.

One of the early published methods dealing with the measurement of the thickness of folded strata was that proposed by Hayes (1908,

p. 37). This method was concerned only with traverses made normal to the strike of the rocks, the variable α being assumed to remain constant throughout the section. Hayes' method involves no geometric assumption regarding the character of the folding but instead is a purely empirical scheme for arriving at a stratigraphic thickness. It was presented as a graphic solution but reduced to a formula reads as follows:

$$t = \frac{s}{2} \cdot \frac{\sin(\delta_1 \pm \sigma) + \sin(\delta_2 \pm \sigma)}{\cos(\delta_2 - \delta_1)} \quad \ldots\ldots\ldots\ldots\ldots\ldots (3)$$

where δ_1 and δ_2 are respectively the lesser and greater values of the dip at any two stations.

An important contribution to this subject was made in 1922 by Hewett (1920, p. 378), who developed a formula for the computation of stratigraphic thickness, under the assumption that with a constant strike the traces of the stratigraphic surfaces in a plane normal to those surfaces should have approximately the form of concentric circles. This assumption also implied the existence of so-called "parallel folds", of which more will later be said. Hewett's formula is

$$t = \frac{h \sin\left(\delta_2 - \frac{\delta_2 - \delta_1}{2} \pm \sigma\right)}{\cos\sigma \cos\left(\frac{\delta_2 - \delta_1}{2}\right)} \quad \ldots\ldots\ldots\ldots\ldots\ldots (4)$$

where δ_1 and δ_2 are defined as above. Hewett also utilized the same assumption in deriving a formula for the true width of an outcrop, where α varied between two stations. The application of the concept of concentric folding was further advanced by Busk (1929, p. 13-34), who developed numerous graphic constructions for securing various types of stratigraphic information. The formulae of Hayes and Hewett, however, together with the graphic methods of Busk, are not only restricted to sections normal to the strike but also imply that the variable α remains constant throughout the measured interval. Recently a paper has been published by Eardley (1938, p. 483-489), in which a graphic method is developed for the measurement of the thickness and curvature of "concentrically folded beds", in a section normal to the axis of a pitching fold. Apparently this method applies mainly to cylindrical folds, where the traces of the beds in a plane normal to the axial line are assumed to have the form of circular arcs. It attempts to reconstruct the traces of the stratigraphic surfaces in this plane but does so only with the aid of the above-mentioned structural assumption. This assumption, however, precludes the application of the method to other types of parallel folding. The method therefore appears to be an empirical one, of limited applicability.

GENERAL CONSIDERATIONS

Before attempting a definition of parallel folds, it is necessary to point out some discrepancies in the current concepts of folded structures. Folds are three-dimensional structures, and, therefore, if three-dimensional geometric descriptions are to be employed, they should properly be referred to surfaces rather than to lines. Thus if a fold is found to have approximately the form of some part of a cylinder, circular or otherwise, such a structure might be called a cylindrical fold, or if it appeared to be a quaquaversal dome or basin, with the general shape of a segment of a sphere or a spheroid, it might be called a spherical or a spheroidal fold. The necessity for this clarification was recently brought to the writer's attention in reading a paper on structural geology, wherein the author, though emphasizing the necessity of a three-dimensional treatment, spoke repeatedly of concentric folds, which in three dimensions should designate a structure where the bedding surfaces have a spherical curvature. But the context showed that the concentric structure was confined to two dimensions, so that a circular cylindrical instead of a spherical fold was really meant. The three-dimensional description of folds is highly desirable, but it does not seem feasible to go beyond the general concepts of elongate or quaquaversal folds, until the geometric forms of the structures have been demonstrated.

From the standpoint of mathematical analysis, folded stratigraphic surfaces are difficult to deal with, because practically nowhere do they assume the shape of simple geometrical figures. Such surfaces can necessarily be represented algebraically only by complex mathematical functions, possibly of a type not amenable to integration, and certainly difficult to derive. The analogous study of such surfaces by purely geometrical methods likewise presents many difficulties, as for example the geometric involution of parallel surfaces and their proper delineation. Both analytically and geometrically, therefore, it seems necessary to deal mainly with plane sections through folded strata. Hence it will be understood that the derivations, figures, and generally the remarks about parallel folds and folding, as given below, refer to two-dimensional profiles, unless otherwise stated.

Even when restricted to two dimensions, however, the term "concentric" is sometimes misused, in that concentric folds or folding are terms utilized by many writers as synonymous with parallel folds or folding. This usage apparently originated with Leith (1923, p. 174), who, in an effort to coin some adjective that was markedly different from "similar", proposed that the term "concentric folds" be used in place of "parallel folds". Geometrically, two curves may be strictly parallel without being concen-

tric, and therefore although all concentric folds are parallel folds, the reverse is not true. Usually the geologist is not in a position to know the true geometric form of a stratigraphic surface, though he may be reasonably sure that the surfaces of certain strata are approximately parallel. In a two-dimensional nomenclature, therefore, it seems best to discard Leith's proposal and in the absence of specific geometric information to speak in general of parallel folds instead of concentric folds.

TYPES OF FOLDS

Folds have been classified broadly by Van Hise, Leith, and others into two general types, known as parallel or competent and similar or incompetent folds. Inasmuch, however, as some structural geologists, like Nevin (1936, p. 50-52), have attempted to define folds other than similar, which are not parallel, it seems best to classify all folds primarily into two classes, parallel and nonparallel. The concept of parallel folds implies principally that the beds, if originally of uniform thickness, remained sensibly so in the process of folding, so that the stratigraphic thickness, measured mathematically at different parts of the section, will be practically uniform. A more rigorous definition of parallel folds is given below. Nonparallel folds, such as similar folds, are conceived to be those in which the beds have been thickened and thinned in the process of folding, a condition that is observable in many sections of incompetent rocks, such as shale. Actually, most folds are probably nonparallel, for it is hard to imagine any considerable degree of folding without some nonuniform thickening or thinning of the beds. But practically there are many sections in which this differential thickening and thinning is so slight that for purposes of stratigraphic measurement the beds may be considered to represent parallel folding. In such sections the use of precise analytical or geometrical methods is believed to be justified.

Where it is evident that differential thickening and thinning is a process of major magnitude, other methods must be used for making stratigraphic measurements. The best method for studying nonparallel folds must necessarily depend upon the stratigraphic data that are available, but in general it would seem that the first approximation to a solution will require the assumption of constant volume during the process of folding. Subsequently, some factor of compression may have to be differentially applied in order to arrive at the true answer. W. W. Rubey (1926, p. 334) has presented one method for measuring the thickness of incompetent beds in nonparallel folds, utilizing the postulate of constant volume. Another method, if the necessary information were available, would be to measure by mechanical integration the area between the traces of two bedding surfaces on a plane normal to such surfaces; and taking for the

length of the arc of the fold some medial line between the two traces, to arrive at a thickness by dividing the integrated area by the length of this arc. It is also possible that this problem could be approached by purely analytical methods, utilizing roulettes or glisettes to duplicate the observed curvature of the traces of bedding surfaces. Yet at best these are two-dimensional methods, and, as nonparallel folds may thin and thicken in three dimensions, such methods are not strictly valid. If, however, the necessary information could be obtained, the method of mechanical integration could be extended into three dimensions by obtaining an integrated volume between two stratigraphic surfaces and by dividing this volume by the area of a medial surface, thus giving a close approximation to the true thickness within some stated areal interval. Analytical methods also might be extended into three dimensions, and in this connection it may be noted that the three-dimensional illustrations of the Bessel function $J_p(x)$ of two real variables x and p, as given by Jahnke and Emde (1933, p. 192, 223), look surprisingly like the conventional idea of Appalachian folding. It is very doubtful, however, whether such three-dimensional analytical methods could be widely utilized, and it is even more doubtful if the results would pay for the mathematical work involved.

PRINCIPLE OF EVOLUTES AND INVOLUTES

Some geometrical concepts must be reviewed in order to formulate a proper definition of parallel folds in two dimensions. When a plane curve rolls without slipping upon a fixed curve, all points on the rolling curve describe a third system of curves, known as roulettes. When the rolling curve happens to be a straight line, the roulettes are known as involutes; and the fixed curve, upon which the straight line rolls, is called the evolute. Obviously, an evolute has an infinite number of involutes. Mechanically, a single involute may readily be described by a pencil anchored to the one end of a thread, which is wrapped around any curve. As the thread is unwrapped, the pencil describes an involute. The geometrical derivation of involutes from a generating evolute is illustrated in Figures 1, 2, and 3.

From the manner of their formation, it is evident that an evolute and its involutes have certain distinctive geometrical properties, as follows:

(1) The rolling straight line is always tangent to the evolute and intersects all the involutes at right angles; and conversely, any normal to an involute is a tangent of the evolute.

(2) Two involutes of the same curve intercept a constant distance on their common normal.

(3) The difference between two radii of curvature of a given curve is equal to the arc of the evolute between the points of contact of the radii with the evolute.

In consequence of the second of these properties, involutes are often called parallel curves, and this is the definition of parallelism in plane curves that is adopted in this paper. Concentric circles are special cases of parallel curves, where the straight line rolls about a point without slipping, to generate its involutes. Here the point is the evolute. Now it is evident that all random plane sections through a parallel fold will not show parallelism of the bedding surfaces, and the number of sections that will show such parallelism depends upon the character of the structure. In a simple structure that has for its axis a straight line, such as a semicylindrical or other elongate fold, the traces of the stratigraphic surfaces on many different planes will show parallelism, but only on certain of these planes will the traces indicate the true thickness and maximum curvature of the beds. Regardless of the curvature of an elongate fold, one such plane is normal to the axis of the fold. If the cross section of the fold is semicircular, an infinite number of other planes can also be drawn through the axis of the fold, showing both parallel traces and true thickness, but not the maximum curvature. And if the cross section of a fold is other than semicircular it can still be shown by the law of the mean that at least one plane may be drawn through the axis of the fold, showing the true thickness. Similarly, in a quaquaversal fold, many plane sections may show parallelism of bedding surfaces, though only certain of these planes may show the true thickness of the beds and the maximum curvature. Finally, in some folds which lack any element of symmetry, no plane section may show such parallelism of the traces of the strata, though one or more curved surfaces intersecting the fold may indeed exhibit such parallelism. But in any study of two-dimensional sections of folds, it seems best to restrict the idea of parallelism to such parallelism as may be observed in plane sections across a fold, though it is evident from the considerations above given that in three dimensions the definition of parallelism might have to be amplified.

All these considerations, then, suggest the following definition. A parallel fold is a fold in which the traces of the bedding surfaces, in one or more plane sections through the structure, will assume approximately the form of involutes, throughout some stated linear interval. Precise measurements of curvature are best made in a plane in which the traces of the stratigraphic surfaces are involutes and in which the maximum curvature is apparent.

APPLICATION OF EVOLUTES AND INVOLUTES

In the introduction of this paper, it was stated that under certain conditions the thickness of a series of folded strata could be satisfactorily measured by substituting the arithmetical means of δ_1 and δ_2, and

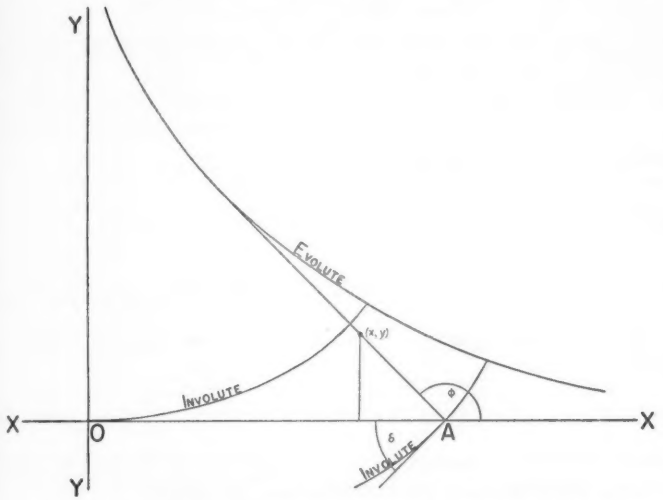

FIGURE 1.—*Sketch illustrating derivation of equation of evolute and involutes, when* $\delta = ka$

of α_1 and α_2, in the homoclinal formula (1). It was also stated that the structural consequences of making such a substitution would also be considered. An analysis of this problem affords at the same time an opportunity to demonstrate the application of evolutes and involutes in structural geology.

Let us suppose that a structure really existed such that $\delta = ka$ where k is a constant, and a is the length of a traverse. Consider, for simplicity, a profile in a plane normal to the strike of the strata, where α is constant. This is shown diagrammatically in Figure 1, where

$a = OA$, the length of a traverse, starting with a zero dip,
$\delta =$ the dip of a stratigraphic surface,
$\phi = arc\ tan\ y'$, for both the evolute and for the normal to the involute.

In order to make the angle δ conform to the numerical magnitudes assigned by the geologist, it is considered to have positive and negative

values, in contradistinction to ϕ, which is taken as always positive. For positive values of δ the construction leads to a syncline; and for negative values of δ the construction leads to an anticline. In the first case, illustrated by Figure 1,

$$a = x + \frac{y}{tan\,(\pi - \phi)} = x - \frac{y}{y'}$$

where x and y are the coordinates of any point located on a normal to the involutes. And $\delta = \phi - \frac{\pi}{2} = arc\,tan\,y' - \frac{\pi}{2} = arc\,ctn\,(-y')$

Therefore $$arc\,ctn\,(-y') = k\left(x - \frac{y}{y'}\right)$$

Hence $$x - \frac{y}{y'} = \frac{1}{k}arc\,ctn\,(-y') \ldots\ldots\ldots\ldots\ldots\ldots (5)$$

$$xy' - y - \frac{1}{k}y'\,arc\,ctn\,(-y') = 0$$

This is a Clairaut equation, so that its solution may be written at once as

$$Cx - y - \frac{C}{k}arc\,ctn\,(-C) = 0 \ldots\ldots\ldots\ldots\ldots\ldots (6)$$

Equation (6) is the equation of any one position of the rolling straight line—*i. e.*, the equation of one normal to the involutes.

The equation of the evolute may be obtained as follows:

Multiply equation (6) by $\frac{1}{C}$, and then differentiate with regard to C:

$$x - \frac{y}{C} - \frac{1}{k}arc\,ctn\,(-C) = 0$$

$$\frac{y}{C^2} - \frac{1}{k\,(1 + C^2)} = 0$$

whence $$C = \pm\left[\frac{ky}{1 - ky}\right]^{\frac{1}{2}}$$

Substituting this value of C in (6), we get

$$y = \pm\left[\frac{ky}{1 - ky}\right]^{\frac{1}{2}} x \mp \frac{1}{k}\left[\frac{ky}{1 - ky}\right]^{\frac{1}{2}} \cdot arc\,ctn\left[\mp\frac{ky}{1 - ky}\right]^{\frac{1}{2}} \ldots\ldots (7)$$

This is the required equation of the evolute.

Since the family of straight lines, represented by equation (5), and the required family of involutes are the orthogonal trajectories of one

another, the differential equation of the involutes may be obtained by substituting $-\frac{1}{y'}$ for y' in equation (5), thus:

$$x + yy' - \frac{1}{k} arc\ ctn\left(\frac{1}{y'}\right) = 0$$

$$x + yy' - \frac{1}{k} arc\ tan\ y' = 0 \ldots\ldots\ldots\ldots\ldots\ldots (8)$$

A parametric solution of this differential equation may be obtained as follows:

Differentiate with regard to y, obtaining

$$\frac{1}{y'} + y' + y\frac{dy'}{dy} - \frac{\frac{1}{k}\cdot\frac{dy'}{dy}}{1+y'^2} = 0$$

Multiplying by $\dfrac{y'\dfrac{dy}{dy'}}{1+y'^2}$, we get

$$\frac{dy}{dy'} + y\left(\frac{y'}{1+y'^2}\right) = \frac{\frac{1}{k}y'}{(1+y'^2)^2}$$

which is linear in y.

The integrating factor is $e^{\int \frac{y'\,dy'}{1+y'^2}} = e^{\log (1+y'^2)^{1/2}} = (1+y'^2)^{1/2}$

Therefore $$y(1+y'^2)^{1/2} = \frac{1}{k}\int \frac{y'\,dy'}{(1+y'^2)^{1/2}} + C$$

$$y(1+y'^2)^{1/2} = -\frac{1}{k}(1+y'^2)^{-1/2} + C$$

whence $$y = \frac{C(1+y'^2)^{1/2} - \frac{1}{k}}{1+y'^2}$$

and $$x = y'\left[\frac{C(1+y'^2)^{1/2} - \frac{1}{k}}{1+y'^2}\right] - \frac{1}{k} arc\ tan\ y'$$

A complete solution of equation (8) is obtained as follows:

From $y = \dfrac{C(1+y'^2)^{1/2} - \frac{1}{k}}{1+y'^2}$, we have

$$y(1+y'^2) - C(1+y'^2)^{1/2} + \frac{1}{k} = 0$$

whence $$(1 + y'^2)^{\frac{1}{2}} = \frac{C \pm \sqrt{C^2 - \frac{4y}{k}}}{2y}$$

$$1 + y'^2 = \frac{C^2 \pm C\sqrt{C^2 - \frac{4y}{k}} - \frac{2y}{k}}{2y^2}$$

$$y' = \left[\frac{C^2 \pm C\sqrt{C^2 - \frac{4y}{k}} - \frac{2y}{k} - 2y^2}{2y^2}\right]^{\frac{1}{2}} = f(y)$$

Substituting $y' = f(y)$ in equation (8), we obtain

$$x + yf(y) - \frac{1}{k} arc\ tan[f(y)] = 0 \ldots\ldots\ldots\ldots\ldots (9)$$

This is the required equation of the involutes. In order to chart each involute, it will be necessary first to assign a fixed value to C, and then to compute $f(y)$ for assigned values of y. Substituting the resulting values of $f(y)$ in equation (8), the corresponding values of x will be obtained.

In general, then, the analytical method of determining the equation of the traces of the stratigraphic surfaces, in a plane normal to those surfaces, and also of determining the equation of the corresponding evolute, is as follows:

(1) Determine the relationship $\delta = f(a)$. In actual practice, a graph may have to be prepared, showing the intercepts (the sites of observations) plotted against the dips at the corresponding stations. Then, by any one of several methods of curve fitting, the algebraic form of $f(a)$ can be obtained.

(2) From $\delta = arc\,ctn\,(\mp y')$, $\delta = f(a)$, and $a = \pm\left(x - \frac{y}{y'}\right)$ determine the differential equation of the normals to the involutes—*i. e.*, $F(x, y, y') = 0$. This will always be a Clairaut equation, so that its solution may be written at once as $G(x, y, C) = 0$.

(3) Differentiate $G(x, y, C) = 0$ with regard to C, obtaining $C = E(x, y)$; substituting $C = E(x, y)$ in $G(x, y, C) = 0$, the equation of the evolute is obtained.

(4) Substitute $-\frac{1}{y'}$ *for* y' *in* $F(x, y, y') = 0$ which will give the differential equation of the family of involutes. The quadrature

of this equation will give the actual equation of the traces of the stratigraphic surfaces in a plane normal to those surfaces—*i. e.*, $H(x, y, c) = 0$.

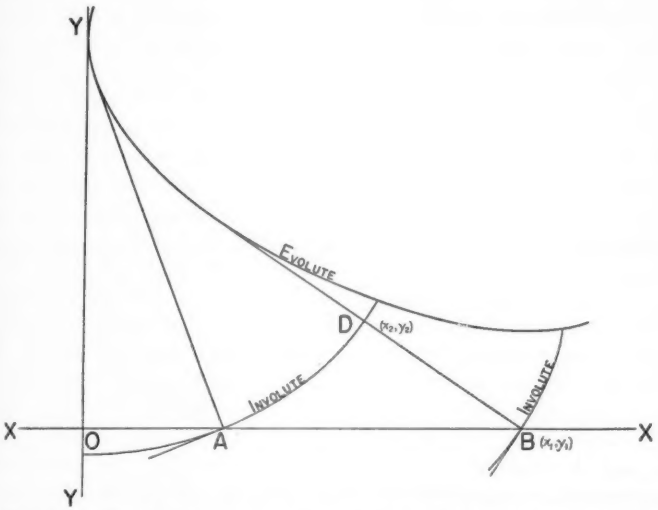

FIGURE 2.—*Sketch illustrating measurement of stratigraphic thickness by the use of an evolute and its involutes*

If a precise measurement of the stratigraphic thickness is also desired by the analytical method of evolutes and involutes, the following method applies:

In Figure 2, let

$OA = a_1$

$OB = a_2$

$AB = (a_2 - a_1) =$ the length of a traverse

$(x_1, y_1) =$ coordinates of point B

$(x_2, y_2) =$ coordinates of the intersection of the involute that passes through A, with the normal that passes through B

$BD =$ the stratigraphic thickness corresponding to the traverse AB

(1) In the equation of the involute that passes through A—that is, $H(x, y, c) = 0$—determine c by utilizing the initial condition that the coordinates of one point on this involute are $(a_1\ 0)$. Substituting this value of c in

$H(x, y, c) = 0$, we get $K = (x, y, a_1) = 0$.

(2) Solve simultaneously $K = (x, y, a_1) = 0$ with the equation of the normal that passes through A—that is, $G(x, y, C_2) = 0$ where $C_2 = \mp ctn\ \delta_2$(at B). This gives the coordinates of the point (x_2, y_2) as functions of a_1 and C_2, both of which are known quantities.

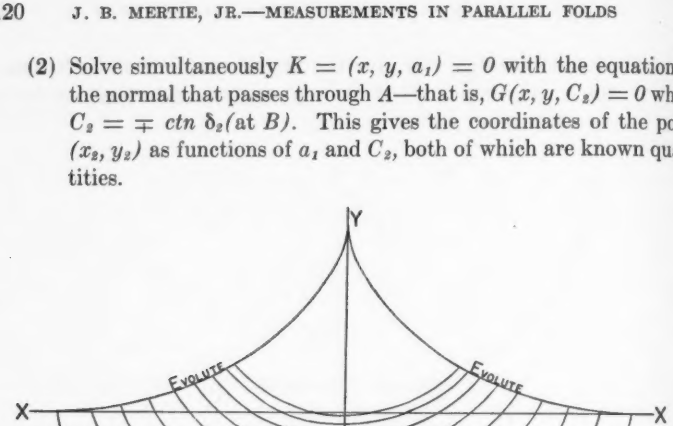

FIGURE 3.—*Diagram showing the type of structure that must exist if* $\delta = ka$

(3) Then in the equation for the length of a straight line, $t = [(x_1 - x_2)^2 + (y_1 - y_2)^2]^{1/2}$, substitute these functions of a_1 and C_2 for (x_2, y_2), and $(a_2, 0)$ for (x_1, y_1). The result gives t, the true stratigraphic thickness for parallel folding.

The evolute, as well as the involute, may also be constructed graphically, and the stratigraphic thickness may likewise be graphically determined. In Figure 3, both the evolute and the involute are laid off graphically, under the assumption that $\delta = ka$. This is done by assuming two Cartesian axes at right angles to one another and by postulating that the dip shall be zero at the origin and $\frac{\pi}{2}$ at some convenient distance, say D, from the origin. From the relationship $\frac{\pi}{2} = kD$, the value of k is found to be $k = \frac{\pi}{2D}$. Then for dips of 10°, 20°, 30° the corresponding values of a are laid off as $a = \frac{2D\delta}{\pi}$. At each of these points—a_1, a_2, a_3—normals are drawn to lines representing the respective dips of the rocks at these points. These normals will not intersect in a point, as in the

conventional concentric construction, but instead will intersect one another in such a manner as to define an envelope of the normals. This envelope, which is easily sketched, is also the evolute of the traces of the stratigraphic surfaces in a plane normal to such surfaces. By the manner of construction it is obvious that the evolute will at some point be tangent to the y axis. Now set pins at close intervals from this point of tangency along the evolute and anchor a string at the point of tangency. If a pencil be placed at the other extremity of the string, and the string be held taut and at the same time be unwound, the pencil will describe an involute; obviously, many pencils at various positions along the string will describe a family of involutes. These involutes define the traces of the stratigraphic surfaces in a plane normal to such surfaces and therefore form a picture of a cross section of the resultant geologic structure, under the assumption $\delta = ka$. More generally, the same process can likewise be carried out for $\delta = f(a)$, when $f(a)$ is any function of a.

The fold shown in Figure 3 is therefore a two-dimensional picture of the type of structure, which is assumed by the geologist to exist when the stratigraphic thickness or some other dimension is computed by substituting the arithmetical means of δ_1 and δ_2, and of α_1 and α_2 in some homoclinal formula. Until the forms of many folds have actually been determined, this assumption may perhaps be as reasonable as any other.

Any conceivable type of parallel fold may thus be uniquely defined in two dimensions by a general relationship $\delta = f(a)$, where $f(a)$ is determined along a line that lies in a plane which is normal to the stratigraphic surfaces and shows the maximum curvature. The function $f(a)$ may obviously assume an infinite number of forms but may always be determined if a sufficient number of observations can be made along or close to a line of traverse, or along any other line in the specified plane. It follows, therefore, that in such a section and with adequate data the traces of the stratigraphic surface, or the family of involutes, together with their evolute, can always be charted geometrically; if the resulting differential equations can be integrated, the algebraic forms of the evolute and its involutes may also be obtained.

The analytical and geometrical derivations above presented obviously apply only to sections which are normal to the stratigraphic surfaces. Similar derivations in sections oblique to these stratigraphic surfaces would necessarily involve the use of more than two variables and would be more difficult to obtain. If it were further desired to extend such derivations into three dimensions in order to learn the geometrical form of a stratigraphic surface, or to obtain an analytical expression therefor, this could theoretically be accomplished by means of two solutions in

plane sections, at least one of which would be oblique to the stratigraphic surfaces, provided the spatial relationships of the two reference planes were known.

Let it not be supposed that these analytical methods are advanced solely for the purpose of deducing the geometrical form of stratigraphic surfaces, or of the traces thereof. In two dimensions, and in certain sections, the form of the traces of stratigraphic surfaces may be learned entirely by geometrical methods, either by the use of involutes or by a piece-by-piece integration of circular arcs. But neither of these geometrical methods yields an analytical expression for the curvature, and the real objective of this exposition is to show how to obtain algebraic equations for the curvature of folds, as a preliminary step in a mathematical analysis of the mechanics of parallel folding or for use in the application of geophysical prospecting methods. If it is possible to determine the algebraic equation of a fold, either in two or three dimensions, it should be possible, at least in some structures, to determine the character and magnitude of the forces that have produced the structure. As an example, consider the deductions that might be made if the involutes, or traces of the stratigraphic surfaces, in some section of a dome should prove to have approximately the geometric form of a catenary. A catenary is the curve assumed by a flexible, inextensible chain, the ends of which are fastened at two points, while the chain hangs at rest under the action of gravity. Would not the geologist be justified, then, in believing that some upward force, acting analogously to gravity but in the reverse direction, had operated to produce the dome? This suggestion is not made with the belief that this method will solve completely the mechanics of folding, but it is certainly one line of approach, which may possibly furnish some useful information in structural problems.

STRATIGRAPHIC MEASUREMENTS

GENERAL CONSIDERATIONS

The limitations imposed in the preceding pages, particularly that limitation which confines the analysis to plane sections in which the traces of the stratigraphic surfaces appear as involutes, do not apply to the approximate measurement of stratigraphic dimensions. On the contrary, stratigraphic measurements in parallel folds may be made in any random plane section by the utilization of empirical methods based upon some geometrical assumption. No apology, moreover, is needed for the use of such empirical methods, because the actual geometrical form of a fold is seldom known. Methods should therefore be employed which make either no assumption regarding the geometrical form of the strati-

graphic surfaces or otherwise some conventional and universally used assumption, such as that of circular curvature for unknown arcs.

The choice of methods to be applied in measuring stratigraphic dimensions in parallel folds will depend upon the available stratigraphic in-

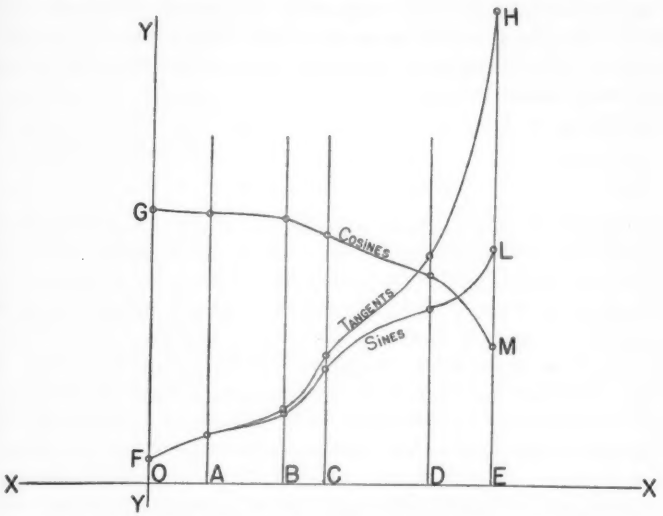

FIGURE 4.—*Diagram showing the derivation of mean trigonometric functions by graphic integration*

formation and also to some extent upon the proclivity of the geologist for graphical or numerical methods. The present paper stresses the use of numerical methods. If several observations of strike and dip can be obtained which lie on or near any random straight line through the bedded sequence, certain empirical but fairly accurate numerical methods can be utilized. But if many observations are available which do not lie on or near a straight line, so that these observations must be utilized in pairs of two, other numerical methods must be found. The case of several observations on or near a straight line will first be considered.

SEVERAL OBSERVATIONS ALONG A LINE

In Figure 4, assume two Cartesian axes at right angles to one another. Consider that an observation of strike and dip is available at the origin and that similar observations are available at *A*, *B*, *C*, *D*, and *E*, corresponding to the distances a_1, a_2, a_3, a_4, and a_5 from the origin. Consider

first only the observations of dip. At *O*, *A*, *B*, *C*, *D*, and *E*, lay off ordinates corresponding to the sines, cosines, and tangents of the dip at these points—say 5°, 10°, 15°, 25°, 40°, and 60°. Now connect these points in smooth curves, utilizing the principle of minimum curvature, and thus obtain sine, cosine, and tangent curves, which correspond to real observations of dip. By mechanical integration, the areas *OFLE*, *OGME*, and *OFHE* can be measured. Call these areas A_s, A_c and A_t. Then for a unit distance *OE*, the mean value of the sines, cosines, and tangents may be written as

$$\overline{\sin \delta} = A_s$$
$$\overline{\cos \delta} = A_c$$
$$\overline{\tan \delta} = A_t$$

Now make similar curves and summations for the successive horizontal angles between the strikes and the direction of the traverse, thus obtaining $\overline{\sin \alpha}$, $\overline{\cos \alpha}$, and $\overline{\tan \alpha}$. The substitution of these mean sines, cosines, and tangents of δ and α, in any homoclinal formula which is a measure of some stratigraphic dimension, such as formulae (1) and (2), will give a close approximation to the magnitude of the desired dimension. The principle involved in this method has already been described by Ickes (1925, p. 451). The same process may be utilized in obtaining mean values of the secant, cosecant, and cotangent, but these functions are seldom required. In this connection, however, one warning is necessary. The cotangent of an angle is the reciprocal of its tangent, but

$$\overline{\tan \delta} \neq \frac{1}{\overline{\operatorname{ctn} \delta}}$$

and the same applies to sines and cosecants, and to cosines and secants. If it is necessary to employ mean secants, mean cosecants, or mean cotangents, these functions must be independently derived from the original data.

In the application of this method, it has been stated that the observations must lie on or near a straight line which penetrates the bedded sequence. If certain observations are not on the straight line, but are close thereto, they may be projected along the line of the strike, so as to appear on the straight line. If two adjacent observations are somewhat farther removed, the traces of the beds may be considered to lie in circular concentric arcs, and the positions of the stations as well as the strike angles may thus be projected onto the straight line by Hewett's (1920, p. 381-382) method. The true values of the dips will not necessarily be the same at the new position as at the original station, except under the special and probably rare condition where the stratigraphic

surfaces have a spherical curvature. But in the absence of specific evidence regarding the three-dimensional curvature of the strata this assumption is warranted. By these and other methods, then, numerous observations may be adjusted to a line, corresponding to a linear traverse. Thereafter the method is applied in the manner shown above.

This method may also be used where a series of observations are considered in pairs. In this case, however, the application consists merely in taking the arithmetical mean between *sin* δ_1 and *sin* δ_2, or between the trigonometric functions of any two initial and terminal observations of δ and α. These mean trigonometric functions are then substituted in any homoclinal formula.

OBSERVATIONS IN PAIRS

It remains to devise a general numerical method for measuring stratigraphic dimensions where many structural observations are available, which cannot be satisfactorily projected onto a straight line, so that the observations have to be used in pairs. One way of solving such problems has already been mentioned in the early part of this paper. The structural consequences of utilizing this method, which is based on the idea of applying the arithmetical means of δ_1 and δ_2 and of α_1 and α_2 in a homoclinal formula, have also been considered. The method of substituting the mean trigonometric functions of δ_1 and δ_2, or of α_1 and α_2, in any homoclinal formula has likewise been mentioned.

Since the publication of Hewett's paper, however, many geologists have been inclined to apply the idea of concentric arcs, in making stratigraphic computations, although this is purely a hypothetical assumption, which cannot in general correspond to the actual geometry of folds. As an illustration, consider a traverse in a plane normal to the axial line of an elongate fold, where the strike is constant and only the angles of dip are being analyzed. If only two observations of dip are available, the normals to the traces of the stratigraphic surfaces will necessarily intersect in a point. One may arbitrarily call this a point of projection, thus interpreting the traces as circular arcs. But if several observations of dip are available, their respective normals will seldom intersect in a common point but instead will intersect in such a manner as to delimit an envelope of the normals, or an evolute, as already explained. Another discrepancy arises if the attempt is made with concentric arcs to show several dips, one of which is vertical. No such construction can be made unless either the station, where $\delta = 90°$, is at infinity, or unless the point of projection is at infinity, in which case all the dips along a traverse will have to be vertical. It is, in fact, for this reason that the tables later presented are not computed beyond 85 degrees.

In three dimensions, where both the dip and strike change from station to station, only two sets of observations are required to show the discrepancy. Thus, in Figure 5, let the plane of the paper lie in the plane of the horizon, and let α_1 and α_2 be strike angles at the two stations

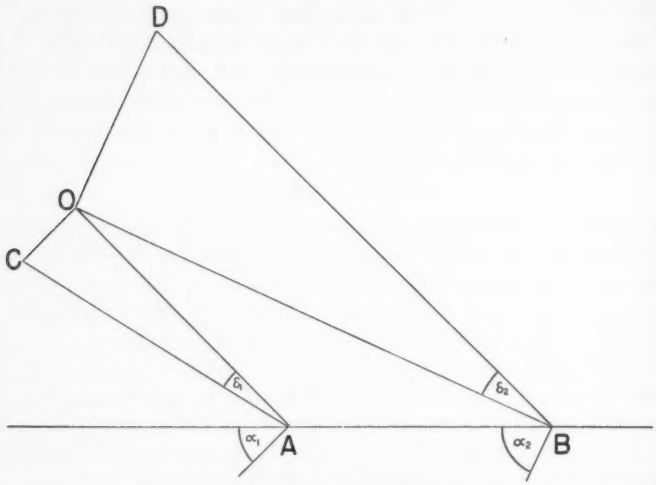

FIGURE 5.—*Sketch showing the fallacy of spherical curvature when applied to random observations of strike and dip*

A and *B*. The corresponding angles of dip are shown by rotating the two planes, which are normal to the stratigraphic surfaces, into the plane of the horizon. If the stratigraphic surfaces could be projected in three dimensions from any one point, the distances *OC* and *OD* should be equal. This is obviously untrue, except in the special case of stratigraphic surfaces having a spherical curvature.

The facts above noted show clearly enough that actual folds seldom have cylindrical or spherical curvature, but the assumption of circular arcs is still warranted in empirical calculations, if its limitations are recognized. One may say, for example, that some kind of curvature exists in stratigraphic surfaces, and, in the absence of any definite knowledge of the geometry of the folds, that the assumption of circular curvature is as good as any other, provided it is not carried beyond its proper limitations. In a sense, it is a matter of curve fitting, where curves of unknown curvature are fitted for short intervals to circular arcs; this is as much justified as other kinds of curve fitting.

In the succeeding pages, an empirical numerical method has been devised for the introduction of the concept of circular arcs into stratigraphic computations, for the most general case where a traverse is oblique to the strike of the rocks, and where both the strike and dip

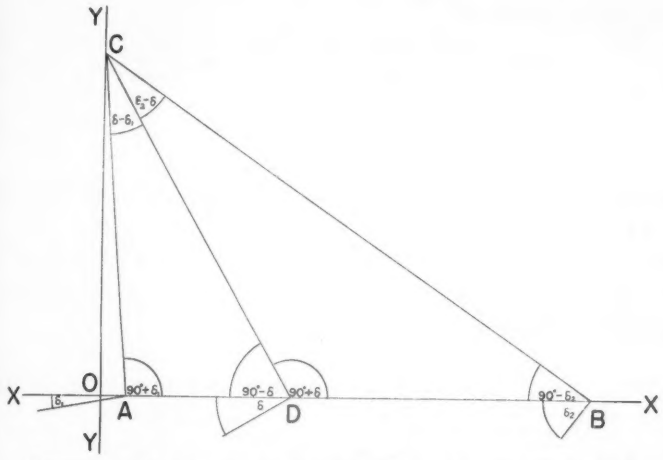

FIGURE 6.—*Sketch illustrating the derivation of* $\delta = arc\ tan\ (ks + g)$

change from station to station. The method, in brief, consists in deriving and evaluating certain integrals, which apply when the dip (or strike) along a line of traverse varies according to an arc tangent law. This empirical postulate, however, does not imply that the stratigraphic surfaces have a spherical curvature, but merely that within certain intervals such a curvature is an approximation to some unknown curvature.

In Figure 6, let A and B be two stations along a line of traverse, normal to the strike of the rocks, and let δ_1 and δ_2 be the dip of the rocks at these two stations. Allow the normals to the stratigraphic surfaces at A and B to be drawn to an intersection C which then is said to define the conventional point of projection. Now let D be some random point along the line AB, and let δ be the dip at D. The distance AB may be treated as a unit distance, so that $AB = 1$, $AD = S$, and $DB = 1 - S$. It is required to find the manner in which δ varies as the point D moves from A to B.

Referring to Figure 6, we have

$$\frac{S\ sin\ (90° \pm \delta_1)}{sin\ (\delta - \delta_1)} = \frac{(1 - S)\ sin\ (90° \mp \delta_2)}{sin\ (\delta_2 - \delta)}$$

$$\frac{\sin(\delta - \delta_1)}{\sin(\delta_2 - \delta)} = \frac{S \sin(90° \pm \delta_1)}{(1 - S)\sin(90° \mp \delta_2)}$$

$$\frac{\sin\delta\cos\delta_1 - \cos\delta\sin\delta_1}{\sin\delta_2\cos\delta - \cos\delta_2\sin\delta} = \frac{S\cos\delta_1}{(1-S)\cos\delta_2}$$

which may be reduced to

$$\tan\delta = \frac{\cos\delta_1\sin\delta_2 - \sin\delta_1\cos\delta_2}{\cos\delta_1\cos\delta_2} \cdot s + \frac{\sin\delta_1}{\cos\delta_1}$$

$$\tan\delta = \frac{\sin(\delta_2 - \delta_1)}{\cos\delta_1\cos\delta_2} \cdot s + \tan\delta_1 \quad \ldots\ldots(10)$$

Now let $\dfrac{\sin(\delta_2 - \delta_1)}{\cos\delta_1\cos\delta_2} = k$ and let $\tan\delta_1 = g$

then $\tan\delta = ks + g \quad \ldots\ldots(11)$

For a unit distance, $AB = 1$, the mean value of the tangents between 1 and 0 is expressed as

$$\overline{\tan\delta} = \int_0^1 (ks + g)\, ds = \frac{k}{2} + g \quad \ldots\ldots(12)$$

Now since $\delta = arc\ tan\,(ks + g)$

$$\sin\delta = \frac{ks + g}{[1 + (ks + g)^2]^{1/2}}$$

And for the same unit distance, $AB = 1$, the mean value of the sines of δ is expressed as

$$\overline{\sin\delta} = \int_0^1 \frac{ks + g}{[1 + (ks + g)^2]^{1/2}}\, ds$$

Let $ks + g = v$

$$ds = \frac{1}{k}\, dv$$

And if $s = 1, v = k + g$; but if $s = 0, v = g$

Therefore $$\int_0^1 \frac{ks + g}{[1 + (ks + g)^2]^{1/2}}\, ds = \frac{1}{k}\int_g^{k+g} \frac{v\,dv}{(1 + v^2)^{1/2}} = \frac{1}{k}\Big[(1 + v^2)^{1/2}\Big]_g^{k+g}$$

$$= \frac{1}{k}\Big[\Big\{1 + (k + g)^2\Big\}^{1/2} - (1 + g)^{1/2}\Big]$$

which reduces to $$\overline{sin\,\delta} = \frac{cos\,\delta_1\,cos\,\delta_2}{sin\,(\delta_2-\delta_1)} \cdot (sec\,\delta_2 - sec\,\delta_1) \ldots\ldots\ldots\ldots (13)$$

Again if $$\delta = arc\,tan\,(ks+g)$$

$$cos\,\delta = \frac{1}{[1+(ks+g)^2]^{\frac{1}{2}}}$$

For the same unit distance, $AB = 1$, the mean value of the cosines of δ between 1 and 0, is expressed as

$$cos\,\delta = \int_0^1 \frac{ds}{[1+(ks+g)^2]^{\frac{1}{2}}}$$

With the same substitutions as before, we have

$$\int_0^1 \frac{ds}{[1+(ks+g)^2]^{\frac{1}{2}}} = \frac{1}{k}\int_g^{k+g} \frac{dv}{(1+v^2)^{\frac{1}{2}}} = \frac{1}{k}\left[log\left\{v+(1+v^2)^{\frac{1}{2}}\right\}\right]_g^{k+g}$$

$$= \frac{1}{k}\left[log\left\{(k+g)+[1+(k+g)^2]^{\frac{1}{2}}\right\} - log\left\{g+(1+g^2)^{\frac{1}{2}}\right\}\right]$$

which reduces to

$$\overline{cos\,\delta} = \frac{cos\,\delta_1\,cos\,\delta_2}{sin\,(\delta_2-\delta_1)} \cdot log\left[\frac{tan\,\delta_2 + sec\,\delta_2}{tan\,\delta_1 + sec\,\delta_1}\right] \ldots\ldots\ldots\ldots (14)$$

Now consider Hewett's formula

$$t = \frac{h\,sin\left(\frac{\delta_2+\delta_1}{2}\right)}{cos\left(\frac{\delta_2-\delta_1}{2}\right)}$$

If h equals unity, and the thickness be calculated for any values of δ_1 and δ_2, the function

$$\frac{sin\left(\frac{\delta_2+\delta_1}{2}\right)}{cos\left(\frac{\delta_2-\delta_1}{2}\right)}$$

may be regarded as equal to a mean value of $sin\,\delta$ in the simple formula $t = h\,sin\,\delta$.

Therefore $$arc\,sin\left[\frac{sin\left(\frac{\delta_2+\delta_1}{2}\right)}{cos\left(\frac{\delta_2-\delta_1}{2}\right)}\right]$$

should be equal to the *arc sin* of the function shown in equation (13). This is found to be true, thus proving the correctness of formula (13).

Three functions have thus been derived, as shown by equations (12), (13), and (14), which give the mean values of the sine, cosine, and tangent of δ, under the empirical postulate of circular arcs. In a similar way, the mean values of the cosecant, secant, and cotangent may be derived; for reference they are also given as follows:

$$\overline{csc\ \delta} = \frac{cos\ \delta_1\ cos\ \delta_2}{sin\ (\delta_2 - \delta_1)}\left[sec\ \delta_2 - sec\ \delta_1 + log\frac{tan\ \delta_1\ (1 + sec\ \delta_2)}{tan\ \delta_2\ (1 + sec\ \delta_1)}\right] \quad (15)$$

$$\overline{sec\ \delta} = \frac{cos\ \delta_1\ cos\ \delta_2}{2\ sin\ (\delta_2 - \delta_1)}\left[tan\ \delta_2\ sec\ \delta_2 - tan\ \delta_1\ sec\ \delta_1 + log\frac{tan\ \delta_2 + sec\ \delta_2}{tan\ \delta_1 + sec\ \delta_1}\right] \quad (16)$$

$$\overline{ctn\ \delta} = \frac{cos\ \delta_1\ cos\ \delta_2}{sin\ (\delta_2 - \delta_1)} \cdot log\left[\frac{tan\ \delta_2}{tan\ \delta_1}\right] \quad (17)$$

In the preceding demonstration, it has been assumed that $\alpha = \frac{\pi}{2}$ which is only a special case of the more general condition that exists when a line of traverse is oblique to the strike of the rocks. But if α is considered to be a variable, a corresponding analysis would have to deal simultaneously with both δ and α, as interdependent quantities, upon the basis of some three-dimensional structural postulate, such as an assumption of spherical curvature. The fitting of small areas of surfaces of unknown curvature to small segments of a spherical surface would not in itself be any more objectionable as an empirical procedure than what has already been done in two dimensions. The resulting functions would be more complex, however, than those which apply in two dimensions and would be much more difficult to derive. The greatest difficulty, however, would arise in their tabulation, because for every computed interval of δ there would correspond a whole series of possible intervals for α. Thus for computed intervals of 5 degrees for δ and α, there would result for the mean sine alone $(153)^2$ tabulated values, and 70,227 values for the mean sine, mean cosine, and mean tangent. This is obviously an impractical procedure. As an empirical method, therefore, it has been assumed that δ and α are independent of one another, and that the same mean values of the sine, cosine, and tangent, which apply to δ may also be applied to α. In other words, it is assumed that another equation similar to equation (11) governs the variation of α. This would be

$$tan\ \alpha = ps + q \quad (18)$$

where

$$p = \frac{sin\ (\alpha_2 - \alpha_1)}{cos\ \alpha_1\ cos\ \alpha_2} \text{ and } q = tan\ \alpha_1$$

Therefore the tabulation below presented applies both to δ and α.

TABLE 1.—*Tabulation of the logarithms of* $\overline{\sin}$, $\overline{\cos}$ *and* $\overline{\tan}$ *of*

	0°	5°	10°	15°	20°	25°	30°	
5°	8.64010–10							
10°	8.94205	9.11622–10						
15°	9.11947	9.24137	9.33575–10					
20°	9.24634	9.33908	9.41464	9.47853–10				
25°	9.34577	9.41965	9.48187	9.53570	9.58326–10			
30°	9.42805	9.48855	9.54075	9.58655	9.62759	9.66480–10		
35°	9.49871	9.54909	9.59323	9.63257	9.66811	9.70060	9.73059–10	
40°	9.56106	9.60342	9.64100	9.67481	9.70562	9.73395	9.76025	9.78
45°	9.61722	9.65296	9.68498	9.71402	9.74063	9.76524	9.78817	9.80
50°	9.66868	9.69878	9.72598	9.75079	9.77365	9.79486	9.81472	9.83
55°	9.71648	9.74169	9.76459	9.78560	9.80503	9.82313	9.84010	9.85
60°	9.76144	9.78229	9.80132	9.81882	9.83508	9.85027	9.86454	9.87
65°	9.80418	9.82112	9.83652	9.85079	9.86407	9.87650	9.88821	9.89
70°	9.84522	9.85841	9.87053	9.88175	9.89221	9.90201	9.91126	9.92
75°	9.88499	9.89470	9.90365	9.91195	9.91971	9.92699	9.93385	9.94
80°	9.92381	9.93022	9.93611	9.94159	9.94673	9.95155	9.95609	9.96
85°	9.96206	9.96523	9.96817	9.97089	9.97344	9.97584	9.97811	9.98

	0°	5°	10°	15°	20°	25°	30°	
5°	9.99936–10							
10°	9.99780	9.99624–10						
15°	9.99493	9.99277	9.98938–10					
20°	9.99085	9.98810	9.98420	9.97919–10				
25°	9.98538	9.98209	9.97765	9.97213	9.96540–10			
30°	9.97738	9.97451	9.96956	9.96350	9.95626	9.94767–10		
35°	9.96959	9.96516	9.95966	9.95310	9.94535	9.93629	9.92573–10	
40°	9.95866	9.95366	9.94763	9.94054	9.93230	9.92275	9.91174	9.89
45°	9.94517	9.93960	9.93303	9.92543	9.91670	9.90671	9.89520	9.88
50°	9.92844	9.92228	9.91518	9.90708	9.89788	9.88745	9.87549	9.86
55°	9.90753	9.90082	9.89315	9.88456	9.87492	9.86409	9.85188	9.83
60°	9.88102	9.87372	9.86556	9.85649	9.84643	9.83523	9.82284	9.80
65°	9.84663	9.83877	9.83012	9.82063	9.81018	9.79865	9.78587	9.77
70°	9.80047	9.79208	9.78299	9.77312	9.76237	9.75060	9.73764	9.72
75°	9.73504	9.72621	9.71675	9.70660	9.69567	9.68379	9.67081	9.65
80°	9.63304	9.62394	9.61430	9.60407	9.59317	9.58144	9.56871	9.55
85°	9.43771	9.42875	9.41943	9.40963	9.39930	9.38828	9.37645	9.36

	0°	5°	10°	15°	20°	25°	30°	
5°	8.64093–10							
10°	8.94529–10	9.12028–10						
15°	9.12703–10	9.24974–10	9.34663–10					
20°	9.26003–10	9.35359–10	9.43160–10	9.49963–10				
25°	9.36764–10	9.44232–10	9.50694–10	9.56482–10	9.61819–10			
30°	9.46041–10	9.52169–10	9.57616–10	9.62598–10	9.67271–10	9.71753–10		
35°	9.54420–10	9.59534–10	9.64174–10	9.68492–10	9.72598–10	9.76586–10	9.80535–10	
40°	9.62279–10	9.66586–10	9.70562–10	9.74314–10	9.77926–10	9.81472–10	9.85017–10	9.88
45°	9.69897–10	9.73541–10	9.76949–10	9.80207–10	9.83377–10	9.86519–10	9.89689–10	9.92
50°	9.77515–10	9.80593–10	9.83508–10	9.86323–10	9.89090–10	9.91857–10	9.94672–10	9.97
55°	9.85374–10	9.87957–10	9.90429–10	9.92843–10	9.95233–10	9.97645–10	0.00119	0.02
60°	9.93753–10	9.95893–10	9.97963–10	0.00000	0.02037	0.04106	0.06247	0.08
65°	0.03030	0.04766	0.06461	0.08143	0.09838	0.11574	0.13383	0.15
70°	0.13791	0.15153	0.16492	0.17832	0.19193	0.20599	0.22074	0.23
75°	0.27091	0.28099	0.29097	0.30103	0.31132	0.32204	0.33339	0.34
80°	0.45265	0.45930	0.46595	0.47271	0.47966	0.48697	0.49475	0.50
85°	0.75701	0.76033	0.76366	0.76709	0.77063	0.77438	0.77841	0.78

$\overline{s}$ and $\overline{tan}$ of δ (or α), under the assumption δ = arc tan (k s + g) and α = arc t

Log $\overline{\sin \delta}$ or Log $\overline{\sin \alpha}$

30°	35°	40°	45°	50°	55°	60°
3059–10						
6025	9.78489–10					
8817	9.80973	9.83007–10				
1472	9.83341	9.85113	9.86805–10			
4010	9.85614	9.87135	9.88592	9.89989–10		
6454	9.87805	9.89090	9.90320	9.91503	9.92643–10	
8821	9.89931	9.90988	9.92002	9.92976	9.93919	9.94834–10
1126	9.92004	9.92841	9.93645	9.94518	9.95165	9.95893
3385	9.94038	9.94661	9.95260	9.95865	9.96393	9.96935
5609	9.96040	9.96454	9.96852	9.97233	9.97603	9.97964
7811	9.98026	9.98231	9.98430	9.98620	9.98804	9.98984

Log $\overline{\cos \delta}$ or Log $\overline{\cos \alpha}$

30°	35°	40°	45°	50°	55°	60°
2573–10						
1174	9.89896–10					
9920	9.88213	9.86705–10				
7549	9.86213	9.84672	9.82891–10			
5188	9.83804	9.82238	9.80435	9.78335–10		
2284	9.80863	9.79273	9.77453	9.75347	9.72872–10	
8587	9.77159	9.75555	9.73732	9.71628	9.69167	9.66219–10
4764	9.72327	9.70722	9.68907	9.66824	9.64395	9.61499
7081	9.65651	9.64064	9.62278	9.60270	9.57874	9.55062
4871	9.55479	9.53944	9.52227	9.50276	9.48021	9.45354
7645	9.36364	9.34955	9.33392	9.31628	9.29598	9.27208

Log $\overline{\tan \delta}$ or Log $\overline{\tan \alpha}$

30°	35°	40°	45°	50°	55°	60°
535–10						
017–10	9.88632–10					
589–10	9.92947–10	9.96357–10				
672–10	9.97589–10	0.00665	0.03976			
119	0.02701	0.05447	0.08424	0.11725		
247	0.08498	0.10910	0.13546	0.16492	0.19868	
383	0.15301	0.17371	0.19652	0.22223	0.25196	0.28742
074	0.23650	0.25365	0.27272	0.29438	0.31909	0.35020
339	0.34560	0.35899	0.37401	0.39111	0.41164	0.43649
475	0.50322	0.51258	0.52318	0.53548	0.55020	0.56840
341	0.78284	0.78779	0.79344	0.80008	0.80815	0.81829

d a = arc tan (p s + q)

60°	65°	70°	75°	80°
9.94834–10				
9.95893	9.96602–10			
9.96935	9.97465	9.97983–10		
9.97964	9.98315	9.98660	9.99000–10	
9.98984	9.99160	9.99331	9.99502	9.99668–10

60°	65°	70°	75°	80°
9.66219–10				
9.61499	9.57948–10			
9.55062	9.51624	9.47240–10		
9.45354	9.42102	9.37967	9.32374–10	
9.27208	9.24306	9.20626	9.15656	9.08179–10

60°	65°	70°	75°	80°
0.28742				
0.35020	0.39097			
0.43649	0.46809	0.51051		
0.56840	0.59194	0.62422	0.67225	
0.81829	0.83169	0.85057	0.87972	0.93200

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The sine, cosine and tangent functions will suffice for most computations, and therefore the common logarithms of the mean sine, mean cosine, and mean tangent have been computed for increments of 5 degrees for δ_1 and δ_2 (or α_1 and α_2). These logarithms are tabulated herewith in Table 1. A more complete table, giving δ_1 and δ_2 (or α_1 and α_2) in increments of one degree, would have been a more desirable contribution, but this would have involved about 25 times as much work, to which the writer was unable to give the necessary time. If the method meets with any general approval, however, facilities may later be found for computing a complete table, not only of sines, cosines, and tangents, but also of cosecants, secants, and cotangents.

An example of a computation of thickness of strata, using these tabulated logarithms, is now given. Let the strike and dip of two bedding planes at stations *A* and *B* be respectively N. 35° E., 10° SE. and N. 70° E., 50° S. Also let the rocks be dipping in a direction opposite to that of the slope of the hillside. And finally let the angle of slope of the hillside be 5°, the direction of the traverse be S. 5° E., and the slope distance be 450 feet. The data, then, are as follows:

$\delta_1 = 10°$ $\quad\quad$ $\delta_2 = 50°$

$\alpha_1 = 40°$ $\quad\quad$ $\alpha_2 = 75°$

$\sigma = 5°$ $\quad\quad$ $s = 450$

From the tables,

$log\ \overline{sin\ \delta} = 9.72598\text{-}10$

$log\ \overline{cos\ \delta} = 9.91518\text{-}10$

$log\ \overline{sin\ \alpha} = 9.94661\text{-}10$

and also

$log\ sin\ \sigma = 8.94030\text{-}10$

$log\ cos\ \sigma = 9.99834\text{-}10$

$log\ 450 = 2.65321$

Substituting these values in the homoclinal equation (1), we obtain the following solution:

$Log\ s$	$= 2.65321$	$Log\ s$	$= 2.65321$
$Log\ \overline{sin\ \alpha}$	$= 9.94661\text{-}10$	$Log\ \overline{cos\ \delta}$	$= 9.91518\text{-}10$
$Log\ \overline{sin\ \delta}$	$= 9.72598\text{-}10$	$Log\ sin\ \sigma$	$= 8.94030\text{-}10$
$Log\ cos\ \sigma$	$= 9.99834\text{-}10$		1.50869
	2.32414	$Anti\text{-}log$	$= 32.3$

$Anti\text{-}log = 210.9$

32.3

Thickness 243 feet

If the dip had been in the same direction as that of the angle of hillside slope, the thickness would have been 210.9 — 32.3 = 179 feet. In a

similar way, these mean trigonometric functions may be employed in homoclinal formula (2) to find the distance or depth to a stratum. But the most important consideration is that by the use of such mean trigonometric functions any formula for any stratigraphic dimension, derived for a homoclinal sequence of rocks, may also be utilized to compute the corresponding dimension in a sequence of parallel folds.

COMPARISON OF METHODS

The concepts of evolutes and involutes has been introduced in this paper principally to demonstrate a new technique for the study of parallel folds. By the application of this principle, the two-dimensional curvature and geometric form of parallel folds may be deduced in certain plane sections normal to the stratigraphic surfaces; and if desired, the corresponding mathematical functions may also be derived through the use of differential equations. It is possible that the same method may also be used in sections oblique to the stratigraphic surface, and conceivably the method may be extended into three dimensions. The information thus obtained should be of material value in any studies of the mechanics of folding and in the application of geophysical prospecting methods.

The method of evolutes and involutes is not recommended for general stratigraphic measurements, unless the line of traverse can be located in a plane in which the traces of the stratigraphic surfaces are involutes which show the maximum curvature; and it should not be used even then, unless a precise measurement, say of thickness, is absolutely indispensable. But if the method of evolutes and involutes can be applied to the measurement of stratigraphic dimensions, a more precise answer will be obtained than by any other method known to the writer. It should also be emphasized that in using this method structural measurements can be made upon the basis of a curvature deduced directly from field data.

For the approximate measurement of stratigraphic dimensions, empirical methods are recommended, because such methods are much less laborious and will yield results that are generally as accurate as the field observations. Many geometrical and numerical methods for this purpose are known, and many others could be devised. In this paper, three numerical methods are given for making stratigraphic measurements, in the most general case where a traverse is oblique to the strike of the rocks, and where both the strikes and dips vary from station to station. These methods are as follows:

(1) The method of substituting in homoclinal formulae the arithmetical means of δ and α. This involves the assumption that

$\delta = k_1 s$ and $\alpha = k_2 s$, where k_1 and k_2 are easily determined constants.

(2) The method of substituting in homoclinal formulae mean trigonometric functions, the values of which are determined by graphic integration. This method involves no assumption whatever regarding the geometry of the folds.

(3) The method of substituting in homoclinal formulae certain definite integrals which result from the assumption that $\delta = arc\ tan\ (ks + g)$ and $\alpha = arc\ tan\ (ps + q)$, where k and g are functions of δ_1 and δ_2 and p and q are functions of α_1 and α_2.

All three of these methods have one point in common—namely, that they utilize homoclinal formulae as the fundamental basis for the stratigraphic computations. One advantage of this feature is that as new homoclinal formulae are developed for special stratigraphic measurements they can be applied at once to strata bent into parallel folds.

No statement is warranted regarding the absolute accuracy of any empirical method, unless it is tested in some folded structure where precise stratigraphic measurements have already been made; and until some precise measurements of some actual structures are made, this comparison is not possible. It is believed, however, that if the necessary data are available the second empirical method affords a close approximation to the true solution; unfortunately, the necessary data are not always to be obtained. The first empirical method has been used for many years, and it is a very convenient one. For small values of δ and large values of α, it yields solutions that differ but little from those obtained with the third empirical method; but for large values of δ and small values of α, it probably introduces errors of appreciable magnitude. The third empirical method utilizes the conventional assumption of circular arcs, and if the second cannot be applied the third is much to be preferred.

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ROCK SERIES IN DIABASE SILLS AT DULUTH, MINNESOTA

BY GEORGE M. SCHWARTZ AND ADOLPH E. SANDBERG

CONTENTS

	Page
Abstract	1136
Introduction	1137
General geology	1138
The Endion sill	1139
Areal and structural relations	1139
Petrography	1140
General statement	1140
Rock number 1	1141
Rock number 2	1143
Rock number 3	1143
Rock number 4	1143
Rock number 5	1144
Rock number 6	1144
Rock number 7	1145
Chemical composition	1145
The Northland sill	1146
Areal and structural relations	1146
Petrography	1148
General statement	1148
Rock number 8	1149
Rock number 9	1149
Rock number 10	1149
Rock number 11	1150
Rock number 12	1150
Chemical composition	1150
Lester River diabase sill	1151
Areal and structural relations	1151
Petrography	1152
General statement	1152
Rock number 13	1153
Rock number 14	1153
Rock number 15	1154
Rock number 16	1155
Rock number 17	1155
Rock number 18	1155
Rock number 19	1156
Rock number 20	1156
Rock number 21	1156
Rock number 22	1157
Chemical composition	1157

Page
Summary of general features of sills 1158
Feldspar problem in red rocks 1160
Origin of the sills 1161
General statement 1161
Red Rock as a separate intrusion 1162
Red Rock by metamorphism of rhyolite 1163
Red Rock by hydrothermal alteration 1164
Red Rock by syntexis 1165
Red Rock by differentiation 1166
Works to which reference is made 1170

ILLUSTRATIONS

Figure Page
1. Sketch map showing location of sills studied and areal geology of the Duluth area 1137
2. Diagram of minerals of the norms of the Endion sill 1142
3. Relative molecular per cent of Na_2O, K_2O, and CaO in rocks of the Duluth sills. 1146
4. Variation in oxides from base of the Endion sill to the rhyolite above 1147
5. Diagram of normative minerals of the five representatives of the Northland sill. 1148
6. Diagram of normative minerals of the seven representatives of the Lester River sill 1154
7. Variation in oxides from base of Lester River sill to the rhyolite above 1158
8. Relative amounts of albite, orthoclase, and anorthite molecules of the norms of the Duluth sills 1160
9. Quartz and orthoclase content of the norms of the Duluth sills 1163
10. Variation diagrams of the Duluth sills and overlying rhyolite flows 1167
11. Alkali-silica ratios of the Duluth sills 1169

Plate Facing page
1. Outcrop map of Endion and Northland sills 1138
2. Outcrop map of Lester River diabase sill 1140
3. Outcrops of Duluth diabase sills 1142
4. Features of the Endion sill 1143
5. Microphotographs of thin sections of Lester River sill rocks 1148
6. Features of the Lester River and Northland sills 1149

ABSTRACT

Above the large gabbro mass in the city of Duluth are several typical Keweenawan sills intrusive into flows of Keweenawan basalt and rhyolite. Three of the sills are large, and each consists of diabase in the lower portion and granite near the top. They were mapped in detail, and samples were selected for petrographic study and chemical analysis.

There is a gradation mineralogically and chemically from the most basic to the most acidic facies. There is, however, a sharp break between the mineral and chemical composition of the granite of the sills and the rhyolite in the roof. The gradational zone within the sill consists mainly of diabase with various amounts of interstitial micropegmatite.

Five possible origins for the rocks in the sill are considered. These are: (1) six separate intrusions of basic and acidic magma, (2) metamorphism of rhyolite by diabase, (3) hydrothermal alteration of the upper portion of the diabase, (4) syntexis, and (5) differentiation of the sill magma.

It is concluded that many factors had some effect, but that differentiation within each sill accounts for the major part of the segregation of the rock in each of the three

sills into two distinct facies. It is difficult to see how crystallization differentiation alone could produce the rock and relations found. Probably a complex process of differentiation involving several processes especially aided by mineralizers might explain the origin of the facies in the sills.

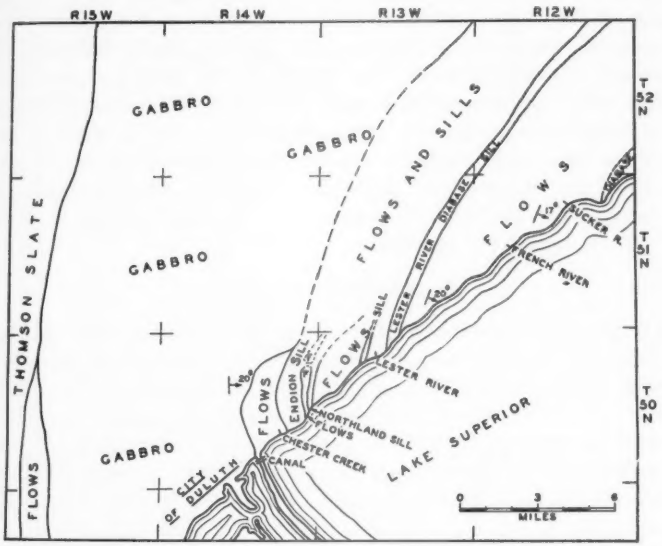


FIGURE 1.—*Sketch map showing location of sills studied and areal geology of the Duluth area*

INTRODUCTION

Intruding the Keweenawan lava flows and sediments immediately above the Duluth gabbro in the city of Duluth is a series of diabase sills. Three of these are thick and although of typical coarse diabase in the lower half they grade to a red granite or "red rock" at the top. In general they show on a small scale the same type of differentiation as the large Duluth gabbro which Grout (1918) described in detail. The Duluth sills are typical of many sills in the Keweenawan areas and also of the Logan sills of the Lake Superior district. They are easily accessible and deserve special attention.

A cross section of these sills is exceptionally well exposed along the shore of Lake Superior which at Duluth cuts across the strike at an angle of about 50 degrees. (*See* Sandberg's (1938) detailed map.)

Some of the flows in which these sills occur are basaltic, and some are rhyolitic, and the question naturally arises as to whether the acidic tops

of the sills are a result of pure differentiation or are, in a measure, a result of the assimilation of felsitic materials. Opportunity to study the sills in detail was afforded when the Minnesota Geological Survey in 1937 assigned to the senior author the task of mapping the Duluth area in detail. Field studies convinced the writers that a combination of field, microscopic, and chemical studies of the sills was highly desirable. The Minnesota Geological Survey has supported the field and microscopic studies, and a grant from the Penrose Bequest of The Geological Society of America made possible the chemical analyses. Aid in the laboratory work was also made available through funds granted by the Graduate School of the University of Minnesota. Professor Frank F. Grout gave frequent advice and read the manuscript. To these sources of support the writers are deeply indebted.

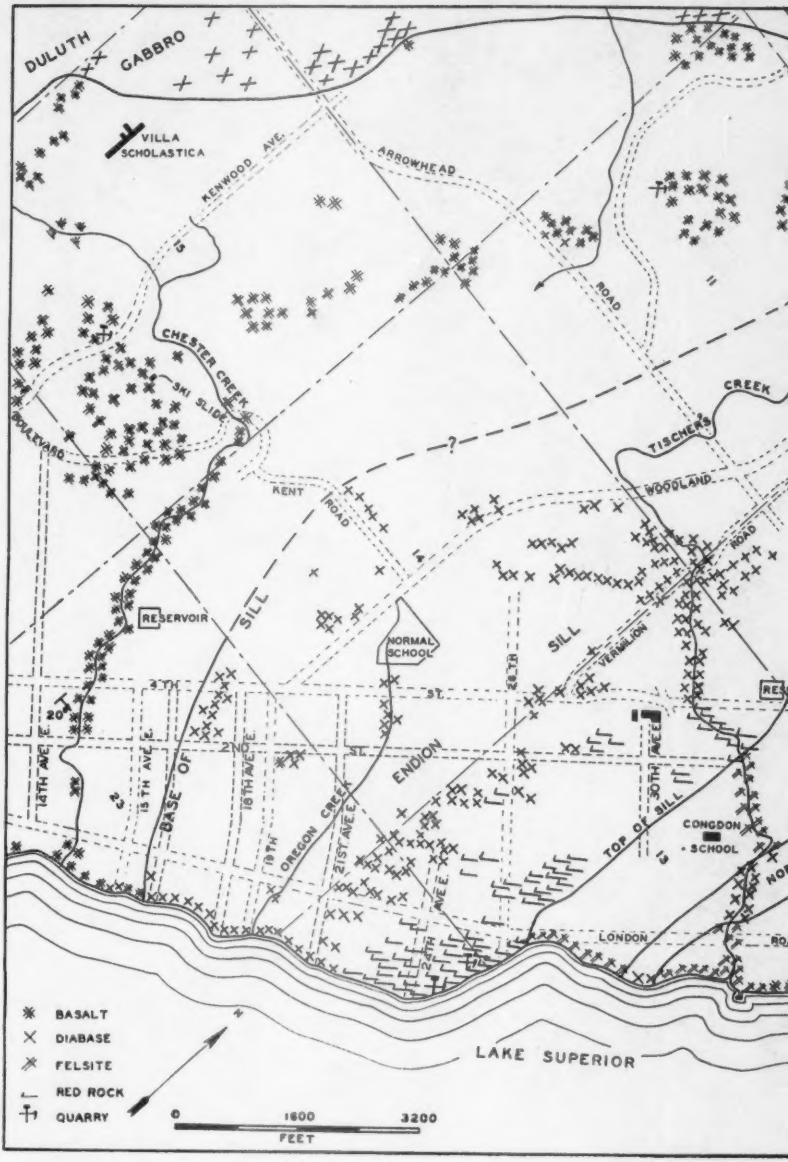
GENERAL GEOLOGY

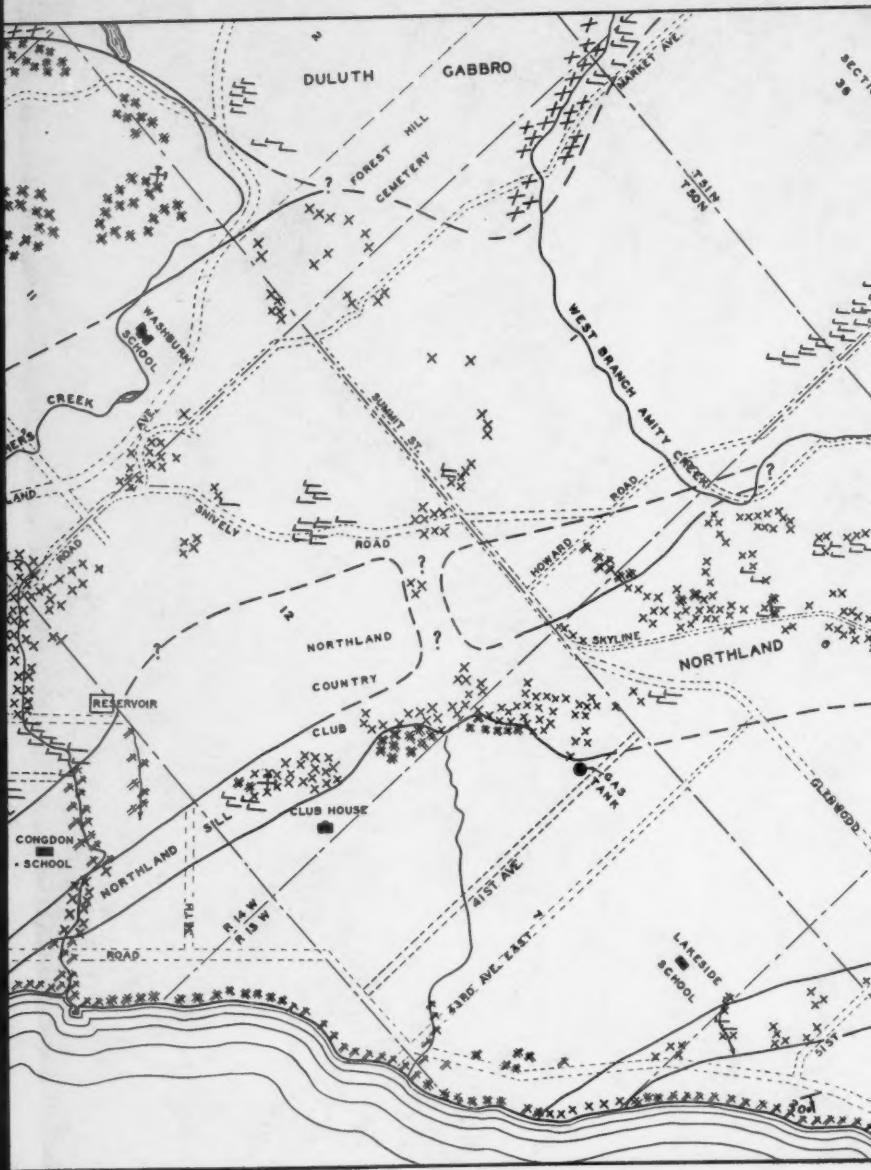
The general areal geology of the region in which the three sills occur is shown by the accompanying map (Fig. 1). At Duluth the Duluth gabbro is intruded into Keweenawan flows which are well exposed both below and above. The base of the gabbro transgresses the flows so that a few miles north, in T. 50 N., R. 15 W., the gabbro intrudes the Huronian (Thomson) slates. There are few exposures of the slate near the gabbro, and north for many miles the base of the gabbro is completely covered by thick glacial deposits. Throughout its entire length the upper contact of the gabbro continues in Keweenawan flows and is well exposed at many places in Duluth and northeastward. The sills are entirely in Keweenawan rocks, mainly basalts and rhyolites.

Since this paper was written a trunk sewer tunnel in Duluth has been driven through several thousand feet of the gabbro and the rocks in the roof. A previously unknown diabase sill was found between the Endion sill and the gabbro. This sill is computed to be 125 feet thick and lies 340 feet stratigraphically above the gabbro. It consists entirely of diabase judging from the exposures in the tunnel.

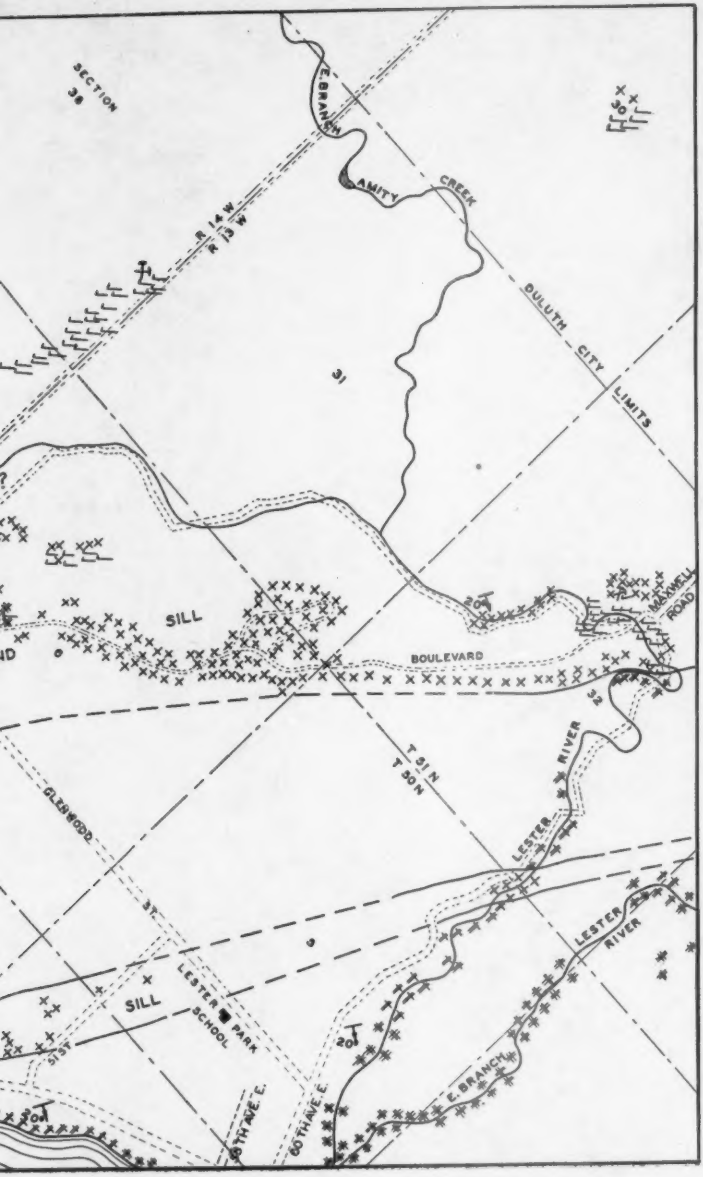
At the shore of Lake Superior[1] the Endion sill (Fig. 1) is separated from the gabbro by 2780 feet of flows and diabase (Sandberg, 1938, p. 805). The gabbro transgresses the flows at such a large angle (Pl. 1) that about 2 miles to the north it is probably in contact with the Endion sill, but unfortunately glacial deposits cover the contact. Farther north outcrops are scarce, but at places the nearest exposures above the gabbro are diabase. It is therefore possible that this sill continues for several miles northward. The Endion sill is not considered an offshoot

[1] In this paper the terms "on shore" or "along shore" refer to Lake Superior.


DULUTH GABBRO
VILLA SCHOLASTICA
KENWOOD AVE.
ARROWHEAD ROAD
CHESTER CREEK
SKI SLIDE
BOULEVARD
TISCHER'S CREEK
WOODLAND ROAD
KENT ROAD
RESERVOIR
SILL
NORMAL SCHOOL
VERMILION ROAD
ENDION
BASE OF SILL
TOP OF SILL
OREGON CREEK
CONGDON SCHOOL
LONDON
LAKE SUPERIOR
14TH AVE. E.
15TH AVE. E.
18TH AVE. E.
19TH AVE. E.
21ST AVE. E.
30TH AVE. E.
2ND ST.
3RD ST.
20TH ST.
BASALT
DIABASE
FELSITE
RED ROCK
QUARRY
0
1600
3200
FEET



OUTCROP MAP OF ENDION AND NORTHLAND SILLS


SECTION 36
E. BRANCH
AMITY CREEK
R. 14 W.
R. 13 W.
DULUTH CITY LIMITS
31
SILL
BOULEVARD
McKNELL ROAD
32
T. 51 N.
T. 50 N.
LESTER RIVER
GLENWOOD
3T.
LESTER PARK
SCHOOL
SILL
31ST
20
30
E. BRANCH
LESTER RIVER
60TH AVE. E.
59TH AVE. E.

of the gabbro because its texture even close to the gabbro is notably fine-grained as compared with that of the very coarse gabbro. An example is the Endion diabase near Forest Hill Cemetery.

The third sill above the gabbro is separated from the Endion sill by about 445 feet of flows at the shore of Lake Superior. This has been called the Northland sill because it forms large rock hills on and near the Northland Country Club. The sill is only 31 feet thick on the shore but thickens rapidly northward and possibly merges with the Endion sill (Fig. 1).

A small diabase sill 132 feet thick is exposed on the shore at the foot of 48th Avenue East between the Northland and Lester River sills (Pl. 1). This sill may be traced for only a short distance northeast by intermittent outcrops but probably is thin all the way and eventually pinches out or disappears under the drift.

The Lester River sill is 3350 feet stratigraphically above the Northland sill and is exposed along shore for slightly over a mile beginning just east of the mouth of Lester River (Pl. 2). The sill may be traced for many miles inland as noted in the detailed description below.

Northeast of the Lester River intrusive there are no sills for several miles along shore, but many others do occur farther along the coast. The most extensive of these have been recently described in detail by Grout and Schwartz (1939). As a rule, thin sills along the north shore of Lake Superior consist only of diabase, whereas many of the larger sills contain both diabase and red rock.[2]

THE ENDION SILL

AREAL AND STRUCTURAL RELATIONS

The flows which the Endion sill intrudes are mainly melaphyres and porphyrites but include thin felsites and are cut by a few narrow dense basalt dikes like that in Figure 1 of Plate 4. At the foot of 12th Avenue East one thick and several thin sandstones are interbedded with the flows.

The Endion sill is about 1300 feet thick and is intruded into or below a rhyolite flow. At its base is about 2 feet of rhyolite overlying a basalt flow (Pl. 3, fig. 1). This is probably a thin portion of the thick rhyolite flow which is exposed above the sill. Sandberg (1938, p. 803) found this flow to be 445 feet thick. The sill is continuously exposed along the shore of Lake Superior from 16th Avenue East to 27th Avenue East in Duluth (Pl. 1). Northward in the city there are numerous exposures of various facies of the sill particularly along Tischer's Creek in Congdon Park, where an excellent section of diabase, granite, and rhyolite is exposed. Farther north on the highland exposures are scattered.

[2] Red rock is a local field term for a granophyric differentiate.

Near Forest Hill Cemetery the Duluth gabbro swings sharply eastward and transgresses the flows which on shore intervene between the gabbro and the Endion sill. It is probable therefore as indicated on Figure 1 and Plate 1 that the flows pinch out and that the gabbro and diabase are in contact although the scarcity of outcrops leaves some doubt. To the northeast in sec. 36, T. 51 N., R. 14 W., and sec. 30, T. 51 N., R. 13 W., isolated granophyr outcrops along the projection of the strike of the Endion sill are probably to be correlated with it.

Outcrops of the rhyolite at the top of the Endion sill are scarce north of Tischer's Creek in Duluth, but the exposures of the Endion and Northland sills are not far apart, and on the Northland Country Club grounds an exposure of diabase may mark a dikelike connection between them. North of the Northland Country Club, the Northland sill trends directly northeast and possibly diverges from the Endion sill, but there are no outcrops over the area between them.

PETROGRAPHY

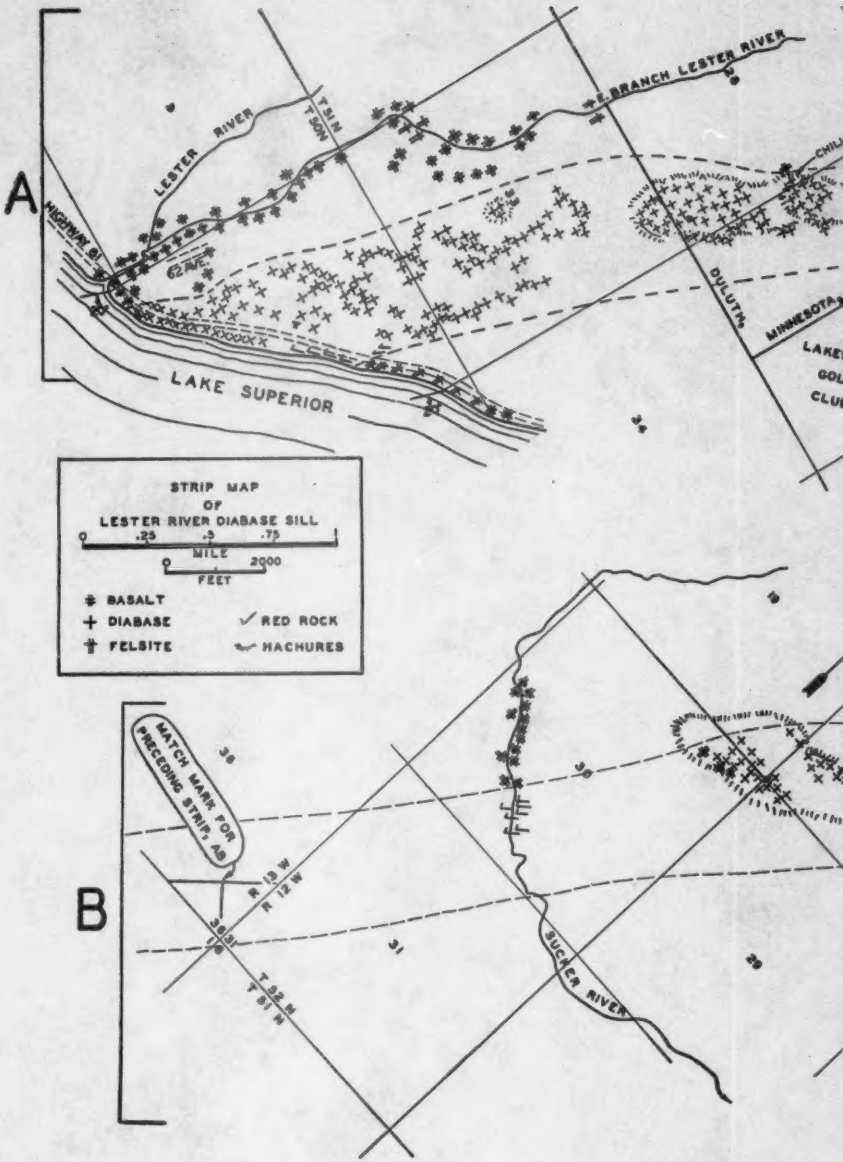
General statement.—The Endion sill is typical of the thicker sills in the Keweenawan flows and Huronian slates of the north shore of Lake Superior. The various facies often grade from one to another, and those selected in the present study for chemical analysis and microscopic study are representative of the various gradations. Megascopic appearance is not always a good guide to significant differences which show up on detailed microscopic and chemical investigation. Six samples of the Endion sill and one of the overlying rhyolite were selected for analysis and miscroscopic study.

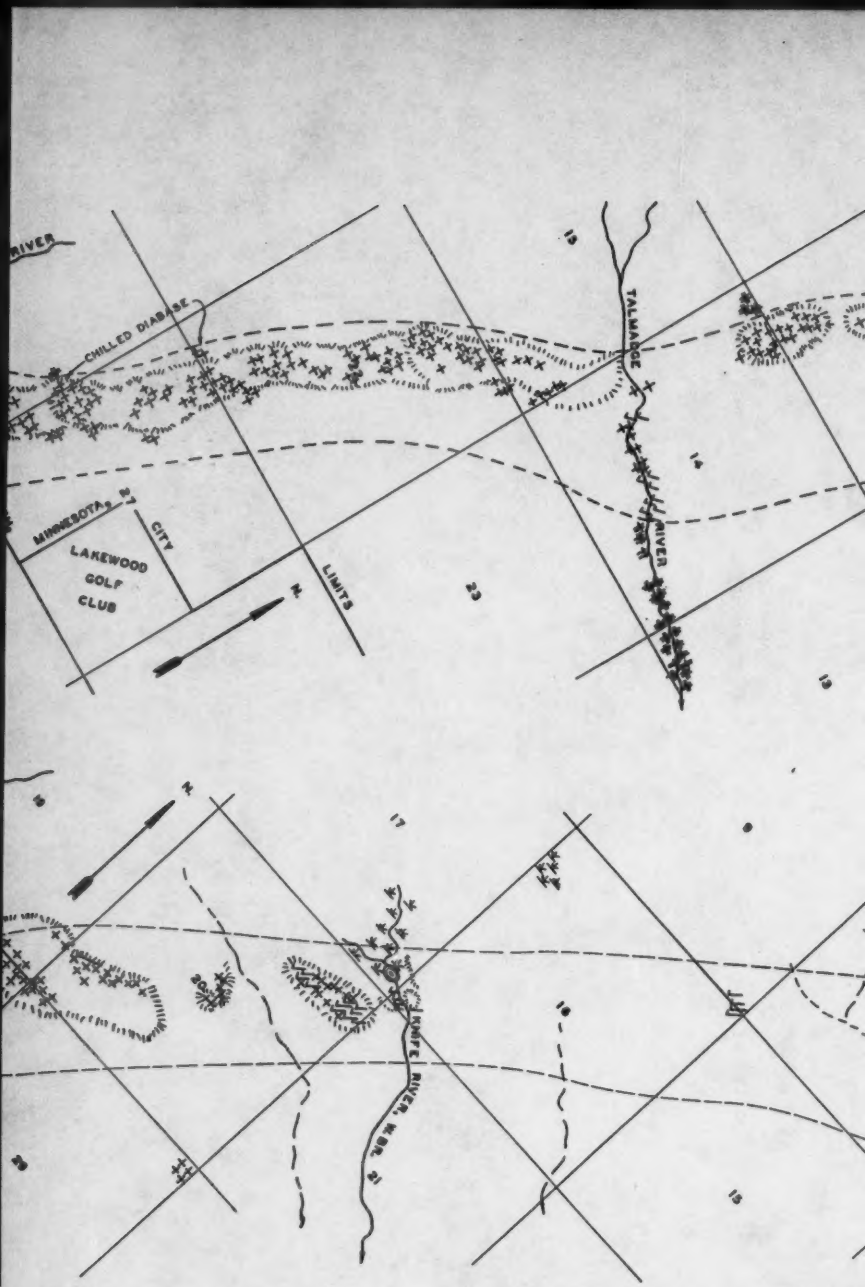
The base of the sill is well exposed on shore at the foot of 16th Avenue East (Pl. 3, fig. 1). At the contact is a chilled basalt containing a few scattered large plagioclase phenocrysts.

This dense zone is not over 6 inches thick and grades upward to the typical diabase. Its thinness suggests the possibility that it is a thin layer of a basalt flow, but detailed examination led to the opinion that it is the chilled border of the sill.

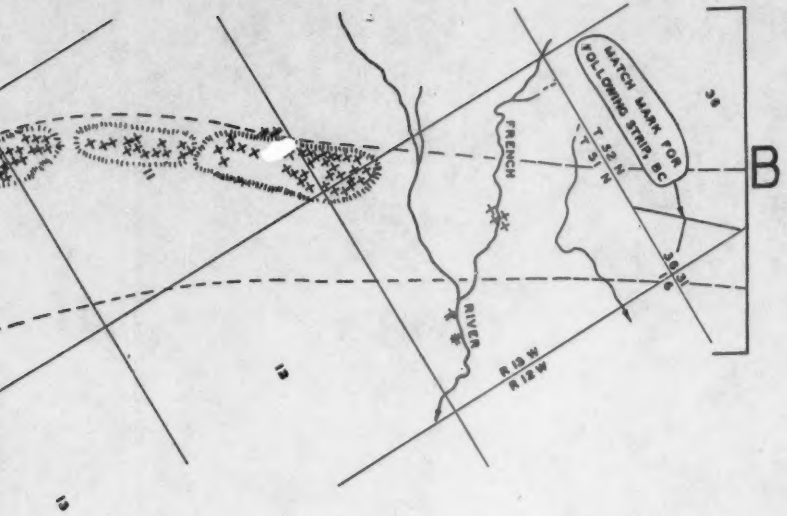
The diabase of the lower part of the sill is black, massive, and ophitic, and blocky jointing is at places rudely columnar. The diabase is very coarsely ophitic (luster mottled) at places, and areas of pyroxene an inch or more across were observed enclosing lath-shaped feldspar crystals (Pl. 4, fig. 3). Pegmatitic diabase patches were also noted in the lower portion of the diabase, and fairly dense stringers of red rock occur in otherwise typical diabase.

The gradation from diabase in the lower part to granitic rock at the top is not uniform, but much of the gradational zone shows a confused


A
LESTER RIVER
T 51N
R 11W
E. BRANCH LESTER RIVER
HIGHWAY 61
62 NE.
DULUTH
MINNESOTA
LAKE SUPERIOR
STRIP MAP
OF
LESTER RIVER DIABASE SILL
.25
.5
.75
MILE
2000
FEET
BASALT
DIABASE
FELSITE
RED ROCK
HACHURES
B
MATCH MARK FOR PRECEDING STRIP, AB
36
R 13 W
R 12 W
T 52 N
T 51 N
SUCKER RIVER
31



OUTCROP MAP OF LESTER RIVER DIABASE SILL


MATCH MARK FOR FOLLOWING STRIP, BC
T 32 N
T 31 N
FRENCH
RIVER
R 10 W
R 12 W
B

FIRE TOWER
T 33 N
T 32 N
C

mixture of diabase and the red acidic facies, probably somewhat later in solidification if not in emplacement.

The presence of fairly dense red rock stringers well down in the sill suggests the existence higher up of acidic liquid at a time when a substantial portion of the sill had crystallized as diabase. Interstitial micropegmatite also probably indicates a later acidic residue. In the vicinity of 21st Avenue East the diabase grades almost imperceptibly into a dark red facies. The gradation to a more typical red rock takes place between 21st Avenue and 22nd Avenue East.

The red rock (granite or granophyr) facies comprises perhaps the upper fourth of the sill. The diabase below it contains small interstitial areas of red feldspar. The red facies ranges from a dark reddish rock which somewhat resembles diabase to a typical red rock or granophyr.

The intermediate or dark red facies forms a considerable portion of the sill and grades to both the more basic and the more acidic varieties as may be readily observed along shore. It is cut by stringers of the most acidic (red) facies (Pl. 4, fig. 2), and small red segregations occur irregularly throughout. Near the rhyolite roof there is a very red zone, representing the most acidic facies in the sill.

Thin sections and chemical analyses show that many dark red facies are rather typical granites. On shore the granite-rhyolite contact is gradational, metamorphism of the rhyolite by the granite having recrystallized the rhyolite so that it is difficult to distinguish the intrusive granite from the rhyolite host. On Tischer's (Congdon) Creek a few blocks northward from the shore, and along the strike, the rhyolite has been less thoroughly recrystallized and the contact is readily located within narrow limits.

Dense basalt dikes cut the diabase and red rock at places but are distinctly later as they have chilled borders (Pl. 4, fig. 1).

The petrographic studies of the rock facies of the sills were made by using thin sections of the same material as was used for chemical analysis. Many other thin sections of rock specimens were also examined but are not described in detail. The rocks are described in order from the base of the sill to the top insofar as this is possible. The general variations in normative mineral composition of the rocks analyzed from the Endion sill are shown graphically in Figure 2.

Rock No. 1.—The sample selected to represent the typical diabase of the sill was collected at the foot of 18th Avenue and below Mile Post 2 of the Duluth, Mesabi, and Iron Range Railroad. This is approximately 600 feet horizontally or 210 feet vertically from the base of the sill.

Possibly another sample closer to the base would have been advisable, but little variation except in grain size was noted and desire to restrict the samples to a reasonable number led to the omission of the basal sample.

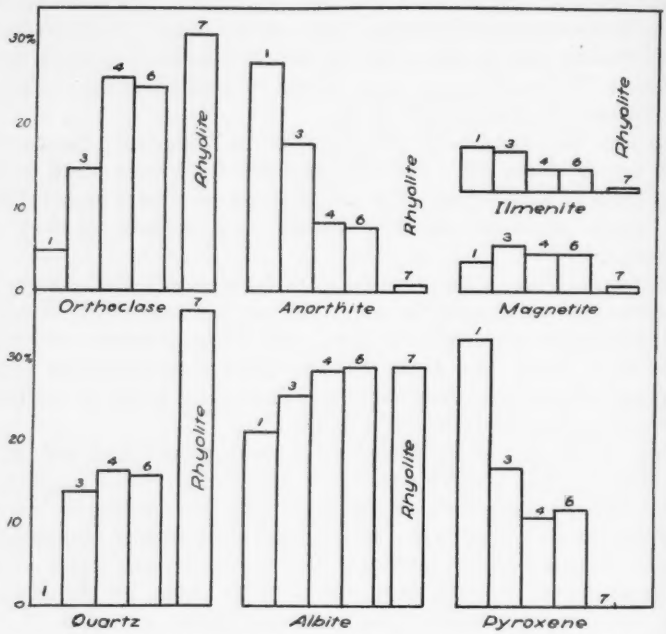


FIGURE 2.—*Diagram of minerals of the norms of the Endion sill*
Arranged in order from the base up.

The rock is massive, black when fresh, and has a fairly coarse diabasic texture with large pyroxenes up to 2.5 mm. across enclosing labradorite crystals (Pl. 4, fig. 3) and rounded inclusions of primary olivine now much altered to an aggregate of uralite, chlorite, serpentine, and talc. Olivine originally made up about 10 per cent of the rock. The labradorite (Ab_{40} An_{60}) remains notably fresh. It is estimated to constitute 65 per cent of the rock. The lath-shaped crystals average about 0.10 mm. by 0.28 mm. A few feldspar crystals are large compared to the averages (0.1 mm. by 0.9 mm.) and result in a faint porphyritic texture. The fresh rock was clearly olivine diabase.

It is noteworthy that this is the only rock in the three sills which has olivine and no quartz in the norm (Table 1). Probably No. 2 would

FIGURE 1. THIN RHYOLITE FLOW (R) OVER BASALT FLOW (B)
Near base of Endion diabase sill (D), 400 feet east of the foot of 15th Avenue East.

FIGURE 2. COLUMNAR JOINTING
In Northland diabase sill. Foot of 30th Avenue East.

FIGURE 3. TYPICAL SHORE LINE OF LESTER RIVER DIABASE
East of the mouth of Lester River.

OUTCROPS OF DULUTH DIABASE SILLS

FIGURE 1. BASALT DIKE IN ENDION DIABASE SILL
Foot of 26th Avenue E.

FIGURE 2. RED STRINGERS IN INTERMEDIATE ACIDIC FACIES
Endion Sill.

FIGURE 3. OPHITIC DIABASE OF ENDION SILL
Foot of 27th Avenue East. X 25.

FEATURES OF THE ENDION SILL

have the same if analyses were complete. The norm is otherwise about as expected of a diabase—that is, low in orthoclase and high in pyroxene.

Another thin section from the main diabase portion of the sill also has a suggestion of porphyritic texture with some euhedral crystals much larger than the average grain size. Magnetite, like augite, has poikilitic plagioclase.

Rock No. 2.—The second sample of the Endion sill was selected from an outcrop on shore 1200 feet horizontally or 420 feet vertically above the base of the sill. This represents the diabase facies of the sill half way between Sample No. 1 and No. 3, which shows some interstitial micropegmatite. Rock No. 2 was selected to check the possibility of variation invisible in the field and shows only a slight difference from the specimen nearer the base. The texture is coarse diabasic with labradorite (0.65 mm. by 5.0 mm.) as the principal primary mineral. Abundant chlorite is pseudomorphous after pyroxene, and antigorite, talc, and fine magnetite have replaced olivine. Pyroxene grains up to 3.2 mm. across enclose small feldspar crystals. Ilmenite and apatite are accessory. This rock is rather fresh looking in hard specimen but has undergone considerable hydrothermal alteration; the labradorite, however, was unaffected except for penetration by chlorite along the cleavage. The unaltered rock was clearly an olivine diabase.

Rock No. 3.—This sample was collected 2500 feet horizontally or 900 feet vertically above the base of the sill, where the visible gradation from diabase to an intermediate type of red rock was first observed.

The texture is granitoid and micrographic, feldspar grains ranging up to 0.5 mm. by 2.5 mm. with pyroxene up to 1.4 mm. across. The primary minerals are mainly plagioclase, orthoclase, and augite, with a small amount (3 per cent) of quartz, and accessory apatite, biotite, magnetite-ilmenite, and sphene. Apatite and magnetite-ilmenite are somewhat abundant for a comparatively acidic rock. Secondary minerals include chlorite, sericite, epidote, hornblende, leucoxene, antigorite, and hematite. Augite has extensively altered to chlorite. The norm has 13.86 per cent quartz, which is much higher than was estimated in thin section. Although much of the feldspar is of low index the norm indicates andesine unless the microcline molecule in plagioclase reduces the index.

Rock No. 4.—The fourth sample, collected in a quarry between 24th and 25th Avenues East near the lake shore, 3300 feet horizontally or 1190 feet vertically from the base of the sill, consists of a medium- to coarse-grained (maximum feldspar grains 0.75 by 1.5 mm.) dark-red granitic rock which has somewhat more ferromagnesian minerals than the

typical red rock and is cut by more acidic stringers (Pl. 4, fig. 2). The rock would usually be classified as syenite in hand specimen.

Plagioclase (mainly albite) and orthoclase are abundant. Some feldspar crystals are zoned, and much micropegmatite is interstitial to the feldspar crystals and to a small extent seems to have attacked them. The normative plagioclase is oligoclase. Quartz is abundant. Much of the feldspar is red with dusty hematite and also contains disseminated sericite. Pyroxene is partly altered to uralite and chlorite. Calcite and serpentine also occur. Feldspar has been penetrated and slightly replaced by chlorite. Magnetite occurs as large grains and crystals, and pyrite was noted replacing one large grain.

This rock illustrates a feature shown by many of the dark granite facies of the sills—namely, an abundance of coarse magnetite-ilmenite. Even when not particularly abundant the magnetite-ilmenite occurs as relatively large grains. The norm shows 6.99 per cent magnetite and ilmenite thus confirming the microscopic observations. (*See also* Figure 2.)

At places the granitic facies is characterized by long needles of hornblende, many of them replaced by an aggregate of alteration products. The individual needles range from 5 to 15 millimeters in length, and sheaflike aggregates of needles up to 3 inches long were observed.

Rock No. 5.—This sample is typical red rock taken from an outcrop on shore opposite 27th Avenue East and within a few feet of the upper contact of the sill with rhyolite. The texture is fine granitoid, porphyritic, and micrographic. Albite and orthoclase constitute about 70 per cent of the rock. Dusty crystals of feldspar (0.3 by 1.1 mm.), some zoned, occur in a groundmass of mainly graphic quartz and feldspar with small amounts of fine-grained pyroxene (0.12 mm.) and magnetite. Alteration is not extensive, but pyroxene has partly altered to chlorite, and small amounts of hematite, sericite, and introduced carbonates also occur. The micrographic quartz and feldspar has attacked the feldspar crystals to a slight extent. The large amount of dusty hematite makes it difficult to determine much of the feldspar. The fact that the soda-potash ratio is 3 to 4 suggests an intermediate feldspar, since only a little albite can be recognized.

Rock No. 6.—The upper contact of the Endion sill is well defined on Tischer's (Congdon) Creek at 2nd Street East about seven blocks (3200 feet) from its mouth.

The sample representing the top of the sill was collected a few feet from the contact with rhyolite. It consists of a dark-red medium-grained (feldspar 0.15 by 0.55 mm.) granitic rock with small blades of hornblende.

TABLE 1.—*The Duluth sills*

ANALYSES

Sample No.	Endion Sill						Flow	Northland Sill					Lester River Sill									Flow
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
SiO_2	47.25	.....	52.42	61.46	.....	61.07	75.48	47.50	58.06	58.01	62.85	63.33	54.86	51.62	50.46	.....	58.88	56.76	61.22	.....	.....	71.12
Al_2O_3	15.00	.....	12.66	13.22	.....	13.66	12.30	12.94	13.09	13.63	13.10	12.99	11.92	11.39	16.90	.....	11.84	12.55	12.30	.....	.....	12.58
Fe_2O_3	2.64	5.97	3.90	3.08	4.53	3.04	2.54	3.94	3.34	4.59	2.71	2.87	5.04	4.94	4.68	3.37	6.82	4.70	9.51	1.96	5.71	5.20
FeO	11.09	9.37	9.55	5.42	2.57	5.54	.36	11.52	7.22	5.19	5.56	5.12	10.32	12.94	6.47	1.75	5.63	8.04	2.93	2.60	.56	.15
MgO	6.52	.....	3.74	2.00	.....	2.48	Tr.	5.62	2.05	3.05	1.40	2.21	1.53	3.00	4.01	.....	2.06	3.57	1.36	.....	.....	.08
CaO	8.40	.....	5.16	2.96	.....	2.36	.14	8.38	4.78	3.63	3.31	2.05	5.84	6.36	8.45	.....	3.63	1.75	1.46	.....	.....	.58
Na_2O	2.52	2.03	3.01	3.33	3.50	3.40	3.43	2.39	3.48	4.23	3.51	3.97	2.96	2.59	2.82	2.60	2.82	2.76	3.24	2.93	2.89	2.85
K_2O	.81	1.17	2.44	4.30	4.59	4.10	5.17	1.07	3.02	3.30	4.02	3.52	2.25	2.03	1.21	5.48	4.07	3.09	3.84	5.75	5.87	6.19
H_2O+	1.63	.....	2.00	1.24	.....	1.79	.24	1.31	1.48	1.02	1.08	1.25	1.68	1.32	1.63	.....	1.18	3.06	1.05	.....	.....	.22
H_2O-	.35	.....	.80	.32	.....	.45	.04	.68	.39	.52	.16	.58	.53	.78	1.28	.....	.58	1.16	.32	.....	.....	.05
TiO_2	2.89	.....	2.66	1.37	.....	1.37	.21	3.74	2.16	2.14	1.59	1.55	1.97	2.26	1.64	.....	1.38	1.62	1.62	.....	.....	.45
P_2O_5	.56	.....	1.14	.40	.....	.48	.02	.69	.70	.41	.40	.37	.66	.54	.42	.....	.41	.55	.54	.....	.....	.03
MnO	.21	.....	.25	.18	.....	.15	.02	.22	.17	.19	.14	.16	.26	.30	.17	.....	.20	.33	.28	.....	.....	.06
CO_2	.....	.....	.....	.53	.....	.20	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.58	.....	.11	.....	.....	.18
Total	99.87	.....	99.73	99.81	.....	100.09	99.95	100.00	99.94	99.91	99.83	99.97	99.82	100.07	100.14	.....	100.08	99.94	99.78	.....	.....	99.74
Sp. Gr. t°/4°	2.965	.....	2.864	2.738	.....	2.718	2.638	2.982	2.818	2.804	2.767	2.728	2.891	2.962	2.818	.....	2.804	2.753	2.797	.....	2.694	2.681

NORMS

Q	.....	.....	13.86	16.20	.....	15.78	35.76	2.64	13.56	10.08	17.58	18.30	15.6	9.60	6.06	.....	17.82	17.22	22.08	.....	.....	29.94
Or	5.00	.....	14.46	25.58	.....	24.46	30.58	6.12	17.79	19.46	23.91	20.57	12.79	12.23	7.23	.....	24.46	18.35	25.58	.....	.....	36.70
Ab	20.96	.....	25.15	28.30	.....	28.82	28.82	20.44	29.34	35.61	29.87	33.54	24.63	22.01	23.58	.....	23.58	23.58	27.25	.....	.....	24.10
An	27.24	.....	17.79	8.34	.....	7.78	.56	20.85	11.12	8.34	7.78	7.23	4.17	13.34	30.02	.....	7.51	5.00	4.17	.....	.....	1.67
Ca	.....	.....	.....	1.20	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.30	.....	.....	.....	.....	.40
Di	9.12	.....	16.60	.46	.....	.....	.....	13.33	6.99	5.75	5.13	.46	17.52	11.79	7.42	.....	3.43	8.71	.....	.....	.....	T.
Hy	23.2	.....	.....	10.31	.....	11.88	.....	19.6	8.62	7.38	6.46	10.02	7.00	18.01	12.09	.....	5.95	8.90	3.40	.....	.....	.20
Ol	1.95	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
Mt	3.71	.....	5.57	4.41	.....	4.41	.70	5.57	4.87	6.73	3.94	4.18	7.19	7.19	6.73	.....	9.98	6.73	5.57	.....	.....	.....
Il	5.47	.....	5.02	2.58	.....	2.58	.46	6.69	4.10	4.10	3.04	2.89	3.65	4.26	3.04	.....	2.74	3.04	3.04	.....	.....	.30
Hm	.....	.....	.....	.....	.....	.....	2.08	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	5.60	.....	.....	5.28
Ap	1.34	.....	2.35	1.01	.....	1.01	.....	1.68	1.68	1.01	1.01	1.01	1.68	1.34	1.01	.....	1.01	1.34	1.34	.....	.....	.....
C	.....	.....	2.14	.....	.....	.71	.92	1.00	.....	.....	.....	.....	3.37	.....	.....	.....	.....	2.75	.82	.....	.....	.51

Endion Sill

No. 1.—Ophitic diabase, shore of Lake Superior at the foot of 18th Ave E., Duluth. 210 feet above base of sill. S. S. Goldich, analyst.

No. 2.—Diabase, shore of Lake Superior at foot of 19th Ave. E., Duluth. 420 feet above base of sill. R. W. Perkin, analyst.

Lester River Sill

No. 13.—Massive, dark, medium-grained diabase. Highway 61, near shore, 1000 feet east of the Lester River bridge, Duluth. 175 feet above base of sill. S. S. Goldich, analyst.

C.....			2.14			.71	.92	1.60					3.37					2.15	.82			.91

Endion Sill

No. 1.—Ophitic diabase, shore of Lake Superior at the foot of 19th Ave E., Duluth. 210 feet above base of sill. S. S. Goldich, analyst.

No. 2.—Diabase, shore of Lake Superior at foot of 19th Ave. E., Duluth. 420 feet above base of sill. R. W. Perlich, analyst.

No. 3.—Diabase at the first sign of gradation to red rock. Shore of Lake Superior opposite 22d Ave. E., Duluth. 900 feet above base of sill. R. W. Perlich, analyst.

No. 4.—Intermediate red rock, McLean quarry near the shore of Lake Superior between 24th and 25th Avenues E., Duluth. 1190 feet above base of sill. R. W. Perlich, analyst.

No. 5.—Red rock a few feet from the contact with rhyolite, shore of Lake Superior opposite 27th Ave. E., Duluth. R. W. Perlich, analyst.

No. 6.—Red rock within a few feet of contact with rhyolite. Tischer's (Congdon) Creek near 2d St. E., Duluth. R. W. Perlich, analyst.

No. 7.—Rhyolite, 50 feet downstream from No. 6. Tischer's (Congdon) Creek near 2d St. E., Duluth. S. S. Goldich, analyst.

Northland Sill

No. 8.—Diabase, shore of Lake Superior, at the foot of 30th Ave. E., Duluth. S. S. Goldich, analyst.

No. 9.—Reddish diabase. Lester River, NW ¼ SW ¼ sec. 32, T. 51 N., R. 13 W. S. S. Goldich, analyst.

No. 10.—Red rock, Skyline Boulevard and 41st Ave. projected, Duluth. S. S. Goldich, analyst.

No. 11.—Intermediate variety of red rock. Lester River, a quarter of a mile west of the center of sec. 33, T. 51 N., R. 15 W. R. W. Perlich, analyst.

No. 12.—Red rock, Lester River above bridge at Maxwell Road, SW ¼ NE ¼ sec. 32, T. 51 N., R. 13 W. S. S. Goldich, analyst.

Lester River Sill

No. 13.—Massive, dark, medium-grained diabase. Highway 61, near shore, 1000 feet east of the Lester River bridge, Duluth. 175 feet above base of sill. S. S. Goldich, analyst.

No. 14.—Dark, massive diabase. Highway 61, 2000 feet northeast of the Lester River bridge, Duluth. 200 feet above base of sill. L. A. Danielson, analyst.

No. 15.—Diabase where red stringers first become abundant, 4600 feet east of the Lester River bridge on the shore of Lake Superior, SW ¼ NE ¼ sec. 4, T. 50 N., R. 13 W. 746 feet above base of sill. S. S. Goldich, analyst.

No. 16.—Red rock of irregular segregations which grade to diabase. Shore of Lake Superior, SW ¼ NE ¼ sec. 4, T. 50 N., R. 13 W. Near No. 15. S. S. Goldich, analyst.

No. 17.—Main facies of red rock. Shore of Lake Superior, SE ¼ NE ¼ sec. 4, T. 50 N., R. 13 W. 825 feet above base of sill. S. S. Goldich, analyst.

No. 18.—Reddish diabase in which No. 19 occurs. Shore of Lake Superior, NE ¼ NE ¼ sec. 4, T. 50 N., R. 13 W. 928 feet above base of sill. S. S. Goldich, analyst.

No. 19.—Red rock along fractures in No. 18. Same location as No. 18. S. S. Goldich, analyst.

No. 20.—Greenish red rock remnant in No. 21. Shore of Lake Superior, NE ¼ NE ¼ sec. 4, T. 50 N., R. 13 W. 928 feet above base of sill. S. S. Goldich, analyst.

No. 21—Red alteration of No. 20. Same location as No. 20. S. S. Goldich, analyst.

No. 22.—Rhyolite above Lester River sill. Mouth of creek on shore of Lake Superior, NE ¼ NE ¼ sec. 34, T. 50 N., R. 13 W. R. W. Perlich, analyst

The texture is granitoid and micrographic. Orthoclase and plagioclase (mainly albite by index, but the norm is oligoclase) constitute 70 per cent of the rock. The feldspar is mainly dusty red with included hematite, and the larger crystals show well-defined zoning. Most of the hornblende has altered to chlorite. Magnetite-ilmenite is fairly abundant. A little introduced carbonate also occurs. There appears to have been some attack on the feldspar crystals by the quartz of the graphic mixtures.

The norm shows 15.78 per cent quartz which is not far from an average for the red rock (granite) of the sills; moreover, there is twice this amount in the rhyolites of the region. Magnetite-ilmenite shows 6.99 per cent in the norm, by coincidence exactly the same as in No. 4. This serves to emphasize the relative abundance of the opaque minerals in the acidic facies of the sills. It should be compared with the decidedly smaller amount in the rhyolite roof as represented by rock No. 7 (Fig. 2).

Rock No. 7.—Rhyolite above the Endion sill was collected 50 feet downstream from the red rock sample on Tischer's (Congdon) Creek. It is a dense red rhyolite with good flow structure. The texture is felsitic and micrographic (feldspar up to 0.02 by 0.07 mm.). Dusty red feldspar and quartz form most of the rock. Magnetite is unevenly distributed, and small amounts of chlorite and sericite also occur. Normative magnetite-ilmenite is 1.16, in marked contrast with 6.99 in the red rock. The rock differs decidedly in structure and texture from the red rock but is mineralogically similar. The amount of quartz is distinctly higher, however. The feldspar is so clouded with hematite that its characteristics are not readily determinable, but it appears uniform and must be anorthoclase according to the norm which shows roughly equal amounts of albite and orthoclase ($Ab_{28.82}$, $Or_{30.58}$).

CHEMICAL COMPOSITION

The composition of the Endion sill is of particular interest because it shows facies similar to those of the Duluth gabbro with which it is so closely associated in its areal occurrence.

The chemical analyses (Table 1) range from a composition typical of a gabbro to that of a granite but not an especially siliceous granite. The range in silica from 47.25 per cent to 61.46 per cent shows that both extremely basic and acidic facies are absent. The range in combined ferrous and ferric oxides from 15.34 to 7.10 per cent with the difference chiefly in ferrous oxide and a corresponding variation in magnesia from 6.52 to 2.00 per cent indicate a smaller amount of ferromagnesian minerals at the top of the sill than at the base. A triangular diagram (Fig. 3) with Na_2O, K_2O, and CaO calculated to total 100 per cent shows the

gradation from high lime and low potash at the base of the sill to rocks containing important amounts of all three oxides in the red rock group. The rhyolite flows above the Endion and Lester River sills (Nos. 7 and 22) in contrast to the red rock show very low lime. It is noteworthy that

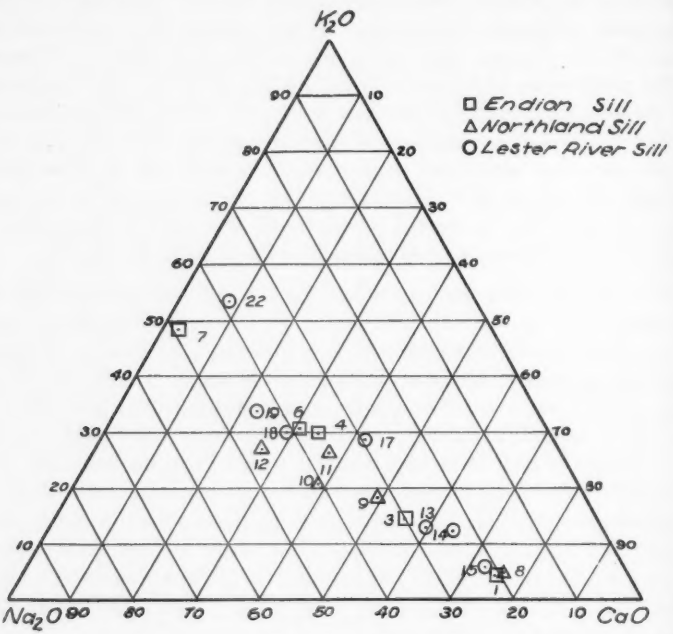


FIGURE 3.—*Relative molecular per cent of Na_2O, K_2O, and CaO in rocks of the Duluth sills*

the rhyolites are in a position by themselves in every diagram which has been made.

The general variation in chemical composition is shown in Figure 4. From bottom to top the increase in silica and potash, the decrease in iron oxides, lime, and magnesia, and the relative constancy of soda and alumina are clearly shown, as is the sharp break between the red rock of the sill and the rhyolite above.

THE NORTHLAND SILL

AREAL AND STRUCTURAL RELATIONS

The Northland sill is much more irregular in plan view than most of the sills in the Duluth region. It is exposed on the shore of Lake

Superior opposite 30th Avenue East and crosses Tischer's Creek in Congdon Park at 33d Avenue East. It is only 31 feet thick on the shore where it consists entirely of diabase. The sill strikes about north-south and dips 17° E. A mile north of the shore, just west of the Club House

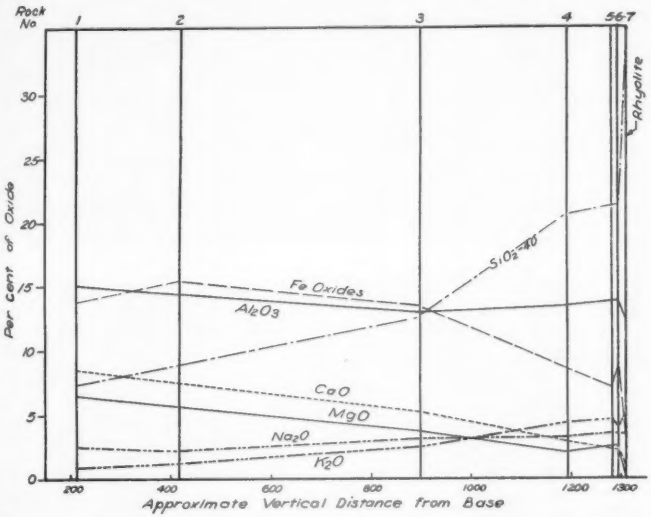


FIGURE 4.—*Variation in oxides from base of the Endion sill to the rhyolite above*

of the Northland Country Club, it is many times as thick and includes a granite facies. The area of outcrop thickens rapidly so that along the Skyline Boulevard in sec. 6, T. 50 N., R. 13 W., there are abundant exposures on a high ridge extending over a width of 1600 feet, and the total width of the sill may be as great as 3000 feet which would indicate a thickness of 1026 feet calculated on the basis of an average dip of 20 degrees.

The relation of the granite to the diabase in this sill is not that of a zone along the top as is usually true of the Endion and Lester River sills, but the granite is found irregularly in the midst of the diabase as shown on Plate 1. Particularly good exposures of the sill occur in the gorge of Lester River where it cuts through the sill. Here, too, granite and diabase occur in confused relation, and diabase is in contact with rhyolite at the top of the sill.

Along the strike of the sill northeast of Lester River are exposures of diabase and granite which are nearly continuous for about a mile.

The ridge is mainly diabase, but granite occurs low on the west side at places. Another exposure occurs half a mile farther north in the SE ¼ NW ¼ sec. 21, T. 51 N., R. 13 W. Still farther northeast for several miles, thick red lake clays and terminal moraine deposits cover the rocks.

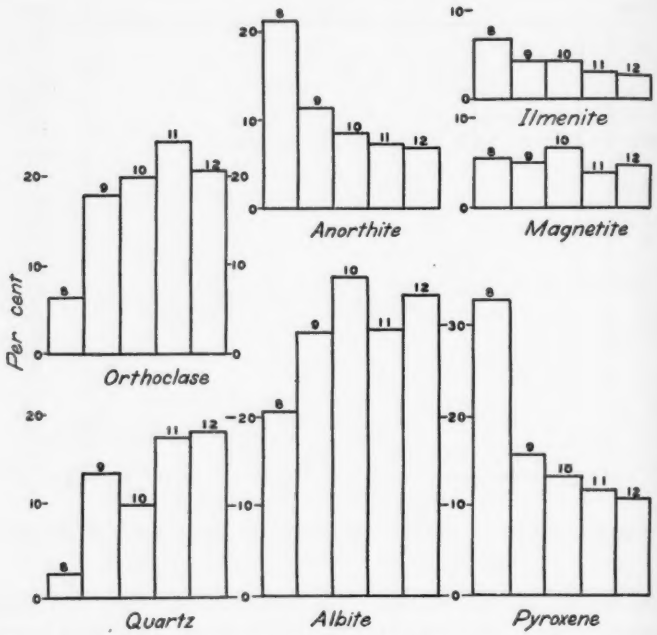


FIGURE 5.—*Diagram of normative minerals of the five representatives of the Northland sill*

Arranged with the most basic at the left.

Several diabase exposures in the northwest part of T. 52 N., R. 12 W., may be correlated with the Northland sill.

PETROGRAPHY

General statement.—The Northland sill as exposed on the shore at 30th Avenue East is dark, fresh-appearing diabase, some showing conspicuous columnar jointing (Pl. 3, fig. 2) and a chilled zone at both bottom and top. There is no granitic facies on shore, but to the north as noted above a confusion of granite in diabase appears as the sill widens greatly. For the petrographic and chemical studies it was not possible to select specimens in order of increasing acidity from the base

FIGURE 1. EUHEDRAL FELDSPAR
Clouded with hematite, some enclosed in quartz.
X 90.

FIGURE 2. LARGE QUARTZ AREA
Partly replaced euhedral feldspar. X 28.

FIGURE 3. CHLORITE (BLACK)
Outlining plagioclase crystals in diabase. X 28.

PHOTOMICROGRAPHS OF THIN SECTIONS OF LESTER RIVER SILL ROCKS

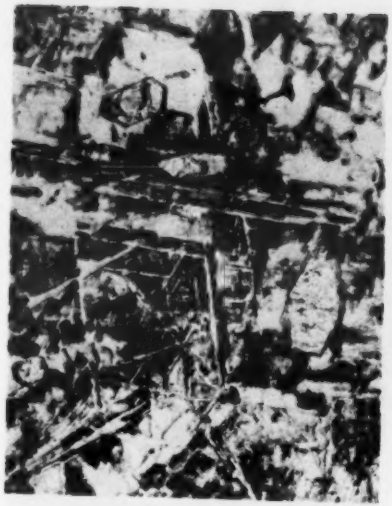


FIGURE 1. LATTICE OF QUARTZ IN FELDSPAR
Typical of the red rock of the Duluth sills. Northland Sill. X 90.

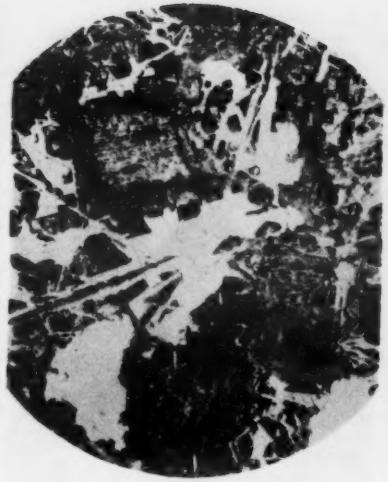


FIGURE 2. ZONED FELDSPAR WITH QUARTZ AREAS AND BLADES
Red rock facies, Lester River Sill. X 21.

FIGURE 3. RED ZONES (DARK)
Along fractures in acidic facies of Lester River Sill.

FEATURES OF THE LESTER RIVER AND NORTHLAND SILLS

of the sill to the top as was done for the Endion and Lester River sills. Samples from five different facies were therefore selected from various positions in the intrusive and these are arranged in Table 1 in order from the most basic to the most acidic, irrespective of position in the sill. The analyses and diagrams presented in the following discussion indicate that the different rocks in the Northland sill follow the general sequence determined for the other sills. (*Compare* Figures 2 and 5.)

Rock No. 8.—The sample representing the normal diabase was taken from the exposure on shore. It is a medium-grained (feldspar 0.2 by 1.2 mm.), massive, black diabase and consists of 60 per cent labradorite, 20 per cent pyroxene, and 15 per cent altered olivine. Magnetite-ilmenite completes the list of primary minerals. Large pyroxenes (2.0 mm.) contain poikilitic feldspar and magnetite. The feldspar remains fresh, but pyroxene is partly altered to tremolite, and olivine has largely altered to iddingsite. Chlorite stringers and veinlets cut the rock.

The norm shows no olivine but 2.64 per cent of quartz. The norm also shows 6.12 per cent of the orthoclase molecule but is otherwise an average type of diabase with high plagioclase (41.29%) and 32.93 per cent pyroxene.

Rock No. 9.—The second facies of the Northland sill, represented by a specimen collected on the West Branch of Lester River in Lester Park (NW ¼ SW ¼ sec. 32, T. 51 N., R. 13 W.), is exposed in the creek bed with a large sloping joint face and in the field was classified as a medium-grained gray to reddish granophyr diabase. The thin section shows that it is much more acidic than is suggested by its appearance in the outcrop or hand specimen. Plagioclase and orthoclase constitute about 65 per cent of the rocks, pyroxene about 18 per cent, and quartz 10 per cent. Many feldspar crystals (0.25 by 2.0 mm.) show well-defined zoning. The pyroxene varies up to about 1 mm. in diameter and is much altered to uralite and chlorite. Micropegmatite is interstitial to feldspar and pyroxene. Considerable sericite has formed in the feldspar, and long blades of quartz are characteristic of the quartz-feldspar intergrowths (Pl. 6, fig. 1). There is very little hematite compared to most of the quartz-bearing rocks of these sills. This accounts for the resemblance to the more basic phases of the sills. Magnetite-ilmenite is relatively abundant (8.97% in the norm), and zoned plagioclase crystals range from andesine to albite.

Rock No. 10.—Another sample chemically much like No. 9 above was collected on Skyline Boulevard at a point about 800 feet west of the center of section 6 in line with 41st Avenue East (Pl. 1). This is closely

associated with a dark porphyry. It is dark red, and the texture is fine granitoid and micrographic and slightly porphyritic. Orthoclase and albite constitute about 65 per cent of the rock, with pyroxene, quartz, and magnetite in order of abundance. Euhedral feldspars are much zoned and clouded with hematite. Feldspar occurs to a small extent in a graphic intergrowth with quartz. Magnetite-ilmenite is unusually abundant as rather large grains (8% estimated in thin section; 10.8% in norm). Pyroxene is partly altered to chlorite and epidote.

Rock No. 11.—Another facies intermediate in appearance between the typical diabase and typical red rock was collected on the West Lester River about a quarter of a mile west of Maxwell Road. This is about 1000 feet northwest of the center of section 32, T. 51 N., R. 13 W. The rock is grayish red and rather dull. The texture is medium granitoid and micrographic. Feldspars are dusty and somewhat reddish and range up to 0.2 by 0.75 mm. Pyroxene is abundant and not much altered. There is considerable interstitial graphic quartz and feldspar. The norm shows 17.58 per cent quartz. Magnetite-ilmenite is moderately abundant, and a few crystals of apatite are accessory. Chlorite is the principal alteration product, but some sericite and hematite also occur.

Rock No. 12.—The most acidic phase of the Northland sill, well exposed at a bridge over West Lester River near the junction of the Lester Park Boulevard and Maxwell Road or about 1000 feet north of the center of section 32, T. 51 N., R. 13 W., is medium granitoid, micrographic, and slightly porphyritic. It contains 70 per cent of plagioclase, ranging from andesine to albite, plus orthoclase and quartz. Lesser minerals are pyroxene, hornblende, magnetite-ilmenite, apatite, and biotite. The magnetite is erratic in distribution. Feldspar occurs as euhedral zoned crystals (0.5 by 1.35 mm.) and with bladelike quartz, mainly as micropegmatite (Pl. 6, fig. 1). Normative quartz constitutes 18.3 per cent. Locally some peculiar euhedral biotite crystals are grouped in quartz. Most of the ferromagnesian minerals have altered to chlorite. Dusty hematite is abundant and masks the characteristics of the feldspar. The similarities and contrasts in the normative minerals of the five facies of the sill are exhibited in Figure 5.

CHEMICAL COMPOSITION

The diabase facies of the Northland sill has a composition essentially similar to that of the diabase of the other two sills. This is in rather sharp contrast to the four other facies of the sill which were analyzed. The four samples analyzed to show various types of the more acidic components of the sill are remarkably alike chemically and indicate

an abrupt break between the acidic and basic facies. The irregular shape, thickness, and the areal relation of the acidic and basic facies of the Northland sill may mean that it is not a single intrusion, but the available outcrops fail to supply evidence on the point. The basic facies is very similar to that of the Endion sill. The four acidic facies were selected for analysis because they appeared different in the field and occurred in different parts of the sill. It is somewhat surprising therefore to find by chemical analysis that the four rocks more acidic than diabase are in fact not radically different. The range in silica, for example, is only 5.32 per cent; alumina less than one per cent. Magnesia ranges from 1.4 to 3.05 per cent which no doubt in part explains the different appearance. Lime shows a similar variation, but soda and potash vary much less. All may properly be classed as granophyr.

LESTER RIVER DIABASE SILL

AREAL AND STRUCTURAL RELATIONS

The Lester River diabase sill was named from its outcrop on Lake Superior just west of the mouth of Lester River near the east end of the city of Duluth (Pl. 2). Its thickness on Lake Superior is given by Sandberg (1938, p. 804) as 963 feet. The strike is about N. 18° E. and the dip 20° SE.

The sill is underlain by a melaphyr flow estimated as 26 feet thick. A narrow covered zone conceals the exact contact. Specimens from an exposure nearest the base are fine to medium diabase and show no evidence of chilling and suggest that they do not represent the actual base. The exposures are practically continuous along shore from this outcrop to the contact with the thick rhyolite flow above. Prominent dip joints near the base strike N. 15° E. and dip 20° SE.

The general extent and exposure of the Lester River sill is well shown by the outcrop map (Pl. 2). It produces a low rocky coast (Pl. 3, fig. 3). It rises gradually with abundant exposures along more or less of a dip slope in and north of sec. 4, T. 50 N., R. 13 W.

In the northeast corner of sec. 33, T. 51 N., R. 13 W., the diabase forms a southward-facing bluff at the end of a ridge which was traced for more than 10 miles northeasterly. It is interrupted by the major stream valleys, and small streams originate on and flow down the dip slope (Sandberg, 1938, p. 807). The ridge shows mainly outcrops of the massive and resistant diabase facies. Some stream crossings have exposed the sill extensively, particularly the acidic facies. It is noteworthy that the strike of the sill, in conformity with that of the flows, swings gradually more easterly to the northeastward. As noted, the strike on Lake

Superior is about N. 18° E. In T. 52 N., R. 12 W., it is close to N. 45° E. (Pl. 2).

Talmadge River exposes both the basic and acidic facies. The usual coarse diabase is well exposed on the ridge on both sides of the valley. The base of the sill is concealed by the red clays deposited in the valley by glacial Lake Duluth. Small exposures of diabase occur about an eighth of a mile downstream from the west line of section 14, T. 51 N., R. 13 W. East of the diabase is a dark granitic facies which grades to a granophyr diabase, which in turn grades downstream into dark granitic facies and thence into the typical red rock exposed near the bridge at the center line of section 14. Closely spaced joints in these exposures strike N. 20° E. and dip 13° SE. The upper contact is also covered, but rhyolite and basalt flows are exposed only a short distance from the granite.

Northeast of Talmadge River the diabase is again well exposed. French River exposes only one outcrop which is normal diabase.

Sucker River, still farther northeast, has extensively exposed basalt flows which appear to lie below the Lester River sill (Pl. 2), but the sill is represented only by granophyr ranging from a rather dark brown type to the typical red facies. This occurs where the lower portion of the sill would be expected.

Northeast of Sucker River the diabase crops out on a series of isolated hills along the strike. A road cut in the diabase exposes a black medium-coarse diabase-gabbro with blocky jointing.

A high knob and the stream valley of the west branch of Knife River also expose the acidic facies. Farther northeastward exposures are scarce, and although all those seen are nearly all granite it is not safe to conclude that the sill is composed entirely of granite. A broad valley floored with glacial lake clay marks exposures beyond section 2, T. 52 N., R. 12 W., and the sill has not been recognized across the valley.

PETROGRAPHY

General statement.—The lower contact of the sill is not exposed, but a somewhat baked amygdaloidal flow is exposed not over 20 feet from the nearest diabase. The diabase is fairly fine-grained but is not chilled, as would be expected at the contact. Exposures are almost continuous along shore across the full width of the sill (Pl. 2; Pl. 3, fig. 3). In the midst of the main diabase are conspicuous mottles about the size of a dime (1.8 cm.). The lower half appears uniform except for small red interstitial patches in the basal portion.

Near the middle of the sill are many one-inch dikelets and segregations of red rock. In the upper portion red zones are related to fractures

in a mixed and greenish rock. There is in reality little difference between the red and green phases except in the amount of oxidation of iron. The acidic zone is not clearly demarcated but is confused from the lowest appearance of red material to the contact of the rhyolite which is not readily located because of recrystallization. Small dense basalt dikes cut the sill, and several clastic dikes were observed in the granitic facies. The clastic dikes contain well-rounded quartz grains as well as fragments of several kinds of rock. The general variations in normative minerals of the rocks analyzed are shown graphically in Figure 6.

Rock No. 13.—The sample selected to represent the basal portion of the Lester River sill was obtained on Highway No. 61 just east of the mouth of Lester River and about 300 feet from the shore of Lake Superior. This is 175 feet horizontally (63 feet vertically) from the base of the sill.

It is a massive black rock with typical diabasic texture except that small interstitial areas of graphic quartz and feldspar also occur. Rather uniform lath-shaped crystals of andesine ranging up to 0.2 by 0.75 mm. form 65 per cent of the rock. Pyroxene and olivine are abundant with pyroxene grains up to 0.9 mm. across enclosing feldspar grains. Magnetite-ilmenite constitutes about 5 per cent. Olivine is much altered to serpentine and to a lesser extent to chlorite. The feldspar is very fresh, and pyroxene is only slightly altered. A few good-sized crystals of apatite occur in feldspar and serpentine. The rock obviously is not a typical diabase because it carries micropegmatite. The chemical analysis shows high alumina and lime. Thin sections of the Endion diabase from near the base show no similar facies. The norm of No. 13 shows no olivine but 15.6 per cent quartz and is unusually low in anorthite (4.17%) for a diabase.

Rock No. 14.—The second specimen from the Lester River sill is more basic. It was collected along the highway 2000 feet northeast of the Lester River bridge. This is about 550 feet horizontally of 200 feet vertically from the base of the sill.

The rock is much like that from near the base, possibly slightly coarser-grained, with feldspars ranging up to 0.3 by 1.0 mm. The rock contains less micropegmatite and more ferromagnesian minerals including augite, olivine, hornblende, and biotite. Augite grains range up to 0.65 mm. in diameter but enclose few feldspar crystals. The feldspar is andesine-labradorite by index and andesine in the norm. Secondary minerals include chlorite, iddingsite, and epidote, but these are less abundant than in the specimen from nearer the base. Olivine, for example, is much less altered.

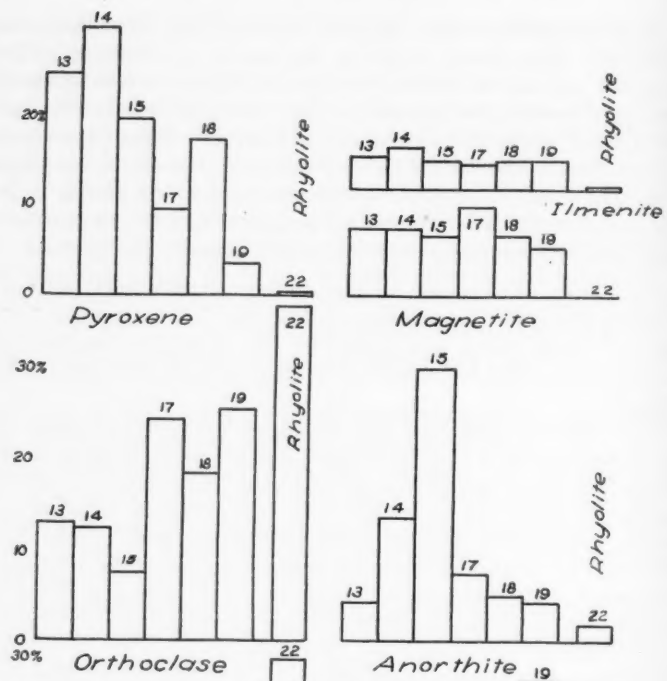


FIGURE 6.—*Diagram of normative minerals of the seven representatives of the Lester River sill*

Arranged in order from the base up.

Rock No. 15.—The third specimen in the Lester River sill series was collected on the shore of Lake Superior near the point where the first red stringers appear. This is opposite Milepost 7 on the Duluth, Mesabi, and Iron Range Railroad, and 746 feet vertically from the base. The

rock is a black coarse-grained diabase (feldspar up to 0.8 by 2.1 mm.) and contains 70 per cent labradorite and 10 per cent pyroxene with grains ranging up to 2.75 mm. Most of the olivine has altered to iddingsite; chlorite is also present. Quartz is subordinate and is mainly interstitial. Accessory minerals include magnetite-ilmenite, biotite, and apatite. A little of the magnetite is found with iddingsite probably liberated during the alteration of olivine.

The norm fails to show olivine as would be expected where quartz is visible in thin section. Normative quartz (9.6%) is much higher than was estimated in the mode.

Rock No. 16.—This sample of the Lester River sill was collected from irregular red segregations which grade into an intermediate type of rock within a few yards of No. 15. The texture is granitoid and micrographic with the larger feldspars ranging up to 0.85 by 1.9 mm. Both plagioclase (albite-oligoclase) and orthoclase are present and form about 65 per cent of the rock. Quartz makes up about 20 per cent, and pyroxene only 7 per cent with grains up to 1.0 mm. across. Quartz poikilitically encloses other minerals (Pl. 5, fig. 2). Much of the feldspar is clouded with dusty hematite and is surrounded by a matrix of quartz. Larger crystals of feldspar are characteristically zoned. Magnetite is scattered throughout as small grains. Chlorite, epidote-zoisite, sericite, and introduced carbonate indicate hydrothermal effects.

Rock No. 17.—This sample represents the main red rock phase. It was collected on shore about 2350 feet horizontally (855 feet vertically) from the base of the sill and near Nos. 15 and 16. It is dark and coarsely granitoid with micrographic quartz and feldspar, the latter ranging up to 0.65 by 2.25 mm. Plagioclase (albite to andesine), orthoclase, and microcline with abundant quartz make up the greater portion of the thin section. Magnetite-ilmenite is surprisingly abundant as coarse grains; these large magnetite grains contain numerous inclusions of chlorite indicating original pyroxene. The norm shows 12.72 per cent magnetite-ilmenite which is higher than any other analyzed facies of the sill (Fig. 6). Epidote and carbonate are other alteration products. Accessory minerals include apatite, zircon, and sphene. The high magnetite-ilmenite content of this main red rock facies probably illustrates concentration of these minerals in the later products of crystallization.

Some of the feldspar areas seem to have been penetrated from the outside by dusty hematite rather than having it uniformly distributed.

Rock No. 18.—The next facies of the Lester River sill occurs about 800 feet along shore northeast of Number 17. It is estimated to be 2550

feet horizontally (928 vertically) above the base and is a dark-greenish to reddish rock (Pl. 6, fig. 3) which superficially appears more like diabase than red rock.

Examination in detail shows a medium to coarse (largest feldspar 0.42 by 1.26 mm.) granitoid and micrographic texture with albite, orthoclase, pyroxene, and quartz and accessory magnetite-ilmenite, apatite, zircon, and a single fluorite crystal. Secondary minerals are chlorite, uralite, sericite, hematite, and serpentine. Chlorite forms secondary rims around laths of feldspar.

Micropegmatite occupies interstices between longer grains. The norm agrees with the conclusion that this is a rather typical granite.

Rock No. 19.—Cutting the greenish rock No. 18 narrow red zones along fractures (Pl. 6, fig. 3) appear very different from their walls in the outcrop but in thin section prove to be little different except that the iron has been oxidized. In hand specimen, the rock is a medium-grained red granite. It consists of oligoclase, orthoclase, and quartz with a fairly high content (8.61% in the norm) of magnetite-ilmenite and accessory apatite. Ferromagnesian minerals are not abundant, and this agrees with the norm which shows only 3.4 per cent pyroxene. Chlorite, sericite, hematite, and carbonate are alteration products. Many euhedral feldspars are zoned and embedded in quartz or graphic quartz and feldspar. Quartz veinlets also penetrate the feldspar. The difference between these two facies is evidently a result mainly of oxidation of ferrous iron to ferric iron thus giving an intense red color to the zones along fractures.

Rock No. 20.—The specimens for Analyses Nos. 20 and 21 were collected on the shore of Lake Superior about 2700 feet horizontally (982 feet vertically) from the base of the sill. They represent greenish-red and red facies of the acidic portion of the sill.

No. 20 represents the greenish type which occurs as remnants in the red. In thin section the greenish-red rock proves to be composed mainly of alkali feldspar and quartz with considerable magnetite-ilmenite. The greenish color is a result of some finely divided chlorite. Other minerals include a little dusty hematite in the feldspar, and carbonate and fluorite.

Rock No. 21.—This brick-red facies as seen in the field is clearly a result of the alteration of the greenish rock. The texture varies from granitoid to micrographic. A few feldspar grains are as large as 1.25 by 2.5 mm. Albite, orthoclase, and quartz with about 5 per cent magnetite compose the greater portion. There is very little chlorite and correspondingly more hematite than in No. 20 and also a little carbonate and fluorite. The feldspar is intensely clouded with this dusty hematite (Pl.

5, fig. 1). Magnetite is partly altered to hematite. Some of the larger quartz areas appear to have replaced feldspar to some extent (Pl. 5, fig. 2), leaving what appear to be angular remnants of red feldspar in quartz.

Rock No. 22.—The rhyolite above the Lester sill is dark red, slightly porphyritic, and consists mainly of quartz and feldspar (albite and orthoclase) with apatite, carbonate, fluorite, zircon, a little magnetite, and much hematite and chlorite as alteration products. The norm is unusual in showing no magnetite but 5.28 per cent hematite. Figure 6 shows the variation in the normative minerals of the specimens described above.

Many thin sections of other specimens from the sill were examined, but only a few show significant differences from those already described.

One thin section shows the contact of a granite stringer in diabase. The diabase is coarse-grained and normal in mineral composition. Much earthy chlorite occurs as an alteration product and outlines the borders of the feldspar crystals (Pl. 5, fig. 3). At the contact quartz corrodes and penetrates andesine, suggesting slight replacement. The granite consists mainly of quartz and orthoclase red with dusty hematite.

A red rock facies collected near Sample No. 17 proves to be an intermediate coarse red rock. The texture is granitoid to micrographic. Feldspar clouded with dusty hematite, quartz, and pyroxene makes most of the rock. Magnetite and apatite occur as accessory minerals, and chlorite and uralite are alteration products of the pyroxene. The noteworthy feature is the prominence of large pyroxene grains in a rock otherwise mainly quartz and feldspar. This thin section shows many features which have been considered typical of the so-called intermediate rock. Other thin sections were examined which contained rather large pyroxene and magnetite grains in a rock otherwise consisting of quartz and feldspar micropegmatite.

CHEMICAL COMPOSITION

For purposes of analysis of the variations within the sill, samples of each distinguishable facies were collected either along shore where wave action has scoured the rocks clean or along the highway where fresh rock cuts have been blasted. Inspection of the analyses (Table 1) shows some significant differences from place to place in the sill. Silica ranges from 50.46 to 61.22 per cent in the samples, but the lowest silica is some distance above the base. The rhyolite immediately above the sill has 71.12 per cent silica or 10 per cent greater than any facies of the sill. Combined FeO and Fe_2O_3 drops from 17.88 per cent in the most basic-appearing sample to 4.56 per cent in one facies. CaO is 8.45 per cent of the facies showing lowest silica and 1.46 per cent of the

acidic facies. Potash varies inversely with lime but not in so great a degree. A more complete idea of the variation in composition between the base and the top of the sill is given by Figure 7. There is more irregularity than in the Endion sill, but the net changes are the same;

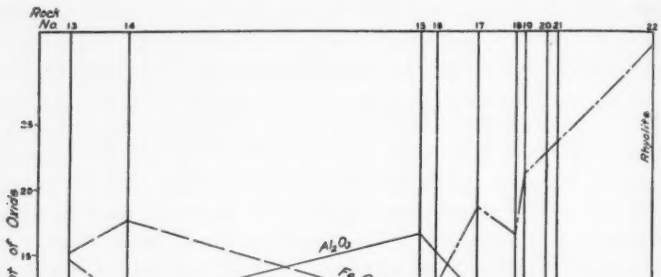


FIGURE 7.—*Variation in oxides from base of Lester River sill to the rhyolite above*

from bottom to top there is an increase in silica and potash, a decrease in iron oxides, lime, and magnesia, and a small net change in alumina and soda.

SUMMARY OF GENERAL FEATURES OF SILLS

The three intrusives described are clearly sills where best exposed on the shore of Lake Superior. Lava flows and sills dip approximately 20 degrees with only minor variations. Inland the Endion and Northland sills become irregular and probably do not maintain a simple sill structure. Irregularity is characteristic of the sills intruded into the Keweenawan flows of the north shore of Lake Superior as has recently been brought out by detailed mapping of a large area by Grout and Schwartz (1939).

The sills are characterized by two types of rock; diabase, ranging from olivine diabase to quartz diabase, and granophyr or red rock as it has often been called. In the Endion and Lester River sills the diabase occurs in the lower part of the sills, and the red rock in the upper

part. The diabase and red rock grade into each other, but it is significant that the gradation is very largely a result of association of the two main types of material rather than a gradation through diorite, syenite, and other types. This situation exists in the Duluth gabbro as was emphasized by Grout (1918). Intermediate rocks exist, and some intermediate facies might readily be classified as syenite in hand specimen, but microscopic examination of many thin sections of all the variations found in the field has failed to reveal a single syenite or diorite specimen.

The diabase of the sills shows no very basic facies, but olivine is present in the most basic facies, and quartz occurs even in some of the olivine-bearing rocks. The feldspar is typically labradorite in the more basic diabase and grades to andesine in those less basic.

The most characteristic red rock consists mainly of red feldspar and quartz, mostly intergrown as micropegmatite. Zoned crystals of feldspar (Pl. 6, fig. 2) occur free from quartz, but quartz rarely forms free grains. It occurs rather as a graphic intergrowth with feldspar or in larger areas of quartz which include euhedral feldspars (Pl. 6, figs. 1 and 2). The norms of the most acidic red rocks show little pyroxene and anorthite but nearly as much magnetite-ilmenite as the diabase. Hornblende which commonly grows as long needlelike crystals is now a mass of alteration products in many of the rocks.

The intermediate facies consists of diabase and granite with rapid gradation on one hand to diabase with only a little interstitial micropegmatite. The acidic facies are characterized by relatively large grains of pyroxene and magnetite-ilmenite with micropegmatite. The amount of the truly intermediate rock is very small compared to the amount of diabase and granophyr.

The feldspars of the acidic portion of the sills are conspicuously dusty with hematite thus giving the whole rock a brick-red color. This dust, however, renders determination difficult. The larger grains are zoned, with more hematite in the outer zones. Analyses of the red rocks show an abundance of soda, potash, and a considerable amount of lime (Fig. 3). In the norm only a portion of the lime is assigned to other minerals so that a considerable percentage of the anorthite molecule is always indicated (Fig. 8). Soda and potash occur in roughly equal amounts so that in sections in which only one feldspar may be recognized anorthoclase is indicated. Intergrowths of feldspar with quartz, and inclusions of euhedral to subhedral crystals of feldspar in quartz, are characteristic.

The differences in the normative minerals in the different facies of the three sills are shown in Figures 2, 5, and 6. Those for the Endion and Lester River sills are arranged from the base up. The three sills

show the same general pattern of decreasing pyroxene and anorthite, relatively constant albite and magnetite-ilmenite, and increasing quartz and orthoclase from the base to the top. The Lester River sill is slightly different; there is a decrease in quartz and orthoclase from the base

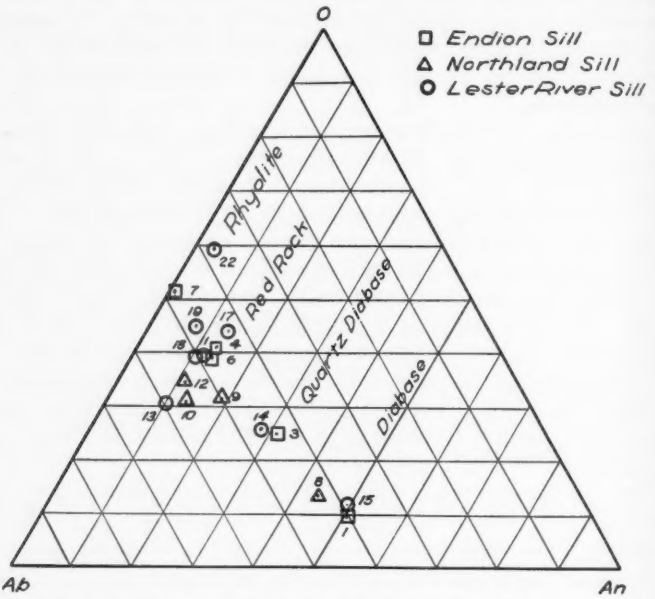


FIGURE 8.—*Relative amounts of albite, orthoclase, and anorthite molecules of the norms of the Duluth sills*

through the first three samples with a corresponding increase in anorthite. The third sample clearly represents a feldspathic facies of the diabase which is not well developed in these sills; however, farther northeast along the shore of Lake Superior such feldspathic facies are developed in some of the large sills like that at Beaver Bay (Grout and Schwartz, 1939). The rhyolites (samples 7 and 22) above the sills consist mainly of quartz and feldspar.

FELDSPAR PROBLEM IN RED ROCKS

The feldspar of the red rocks of the Keweenawan are characteristically red with dusty inclusions of hematite, which renders the exact identification by the common optical methods practically impossible. Tests with index oils on the least dusty material show a range of indices,

but most of the feldspar has a low index, suggesting albite. Many crystals are zoned (Pl. 6, fig. 2), and the outer zone is almost invariably highly clouded with the hematite. Most of the feldspar in the red rock of the Duluth sills shows very little twinning, whereas that in the diabase shows the usual good albite twinning. Many of these features have recently been described for red rocks in sills from farther down the Minnesota coast of Lake Superior (Grout and Schwartz, 1939, p. 42).

In the present study the abundant chemical data available have permitted certain checks on the microscopic work. Figure 3 shows the relative amounts of soda, potash, and lime in the sills. The red rocks as a group have about 1 per cent less lime than potash or soda. In Figure 8 the relative amounts of the three feldspar molecules in the norms are plotted. Inasmuch as considerable lime is assigned to the pyroxenes of the norm this diagram indicates that anorthite is less important than the alkali feldspar molecules, averaging perhaps 14 per cent in the red rocks as a group. Albite and orthoclase molecules are present in roughly equal proportions. In many thin sections the feldspar appears to be of a single species, and efforts to determine this species by microscopic methods have not met with much success because of the dusty red hematite and the lack of twinning. Inasmuch as most slides show a single feldspar of low index, it is probable that this is potash oligoclase of the series designated by Winchell (1927, p. 281) as unstable at low temperatures.

The dusty hematite inclusions are so abundant that it seems probable that iron oxide was in solution at the time of crystallization but was ex-solved during cooling.

ORIGIN OF THE SILLS

GENERAL STATEMENT

The close lithologic and spatial relation of the sills to the Duluth gabbro has been emphasized in the descriptions. It is also significant that the three sills are closely related to the rhyolite flows interbedded with the basalt flows of the region (Sandberg, 1938). At the top the Endion sill is in contact with a rhyolite flow 445 feet thick. The Northland sill intrudes rhyolite, and the Lester River sill is overlain by a 254-foot rhyolite bed. All these rocks are usually dated as Middle Keweenawan.

This close association of the sills with acidic flows raises the question of the relation of the granophyr top of the sills to the intruded rhyolite flows. These are not the only sills in the Keweenawan of the north shore of Lake Superior which show red rock facies. (Grout and Schwartz, 1933; 1939). Other sills on the north shore which do not intrude acidic

rocks nevertheless have granophyr or granite in the sills. It is therefore necessary to look for evidence other than association to determine the origin of the different rock facies within these three sills. The possible modes of origin may be listed as follows:

(1) The red rock is an intrusion distinct from the diabase.

(2) Formation of the red rock facies by metamorphism of the rhyolite to a more granitoid texture.

(3) It has recently been suggested that some red rocks with diabase have been formed by hydrothermal alteration.

(4) Formation of the red rock facies by syntexis—that is, by assimilation of the rhyolite which may be followed by differentiation.

(5) Formation of the different rock facies mainly by differentiation within the sill between the time of intrusion and of complete solidification.

RED ROCK AS A SEPARATE INTRUSION

Granitic intrusion without basic facies is so common that it is logical to inquire whether the granophyr of these sills represents a later intrusion. Bastin (1938) has recently suggested this possibility for the Pigeon Point sill at the extreme eastern end of the Minnesota coast. His evidence for such an origin for the Pigeon Point granite was not conclusive.

As emphasized in the descriptions of the Duluth sills they are remarkably well exposed in cross section along the coast of Lake Superior which cuts across the sills at angles varying from 30 to almost 90 degrees to the strike. There is accordingly an excellent opportunity to see the relation of the red rock facies to the diabase. That the usual features to be expected at the contact of a later granite with diabase do not exist has seemed clear to the writers and to other geologists who have gone over the ground. The smaller occurrences of red rock in diabase are discontinuous which indicates that they formed as a part of the sill, as they could scarcely have been injected into solid diabase. Moreover, in the Duluth area at least, detailed mapping by the writers has not revealed red rock anywhere except adjacent to or within diabase or gabbro intrusives of considerable size. Furthermore, there is nothing about the diabase sills that would make them preferred places for the intrusion of later magmas. The close, if not exclusive, association of red rock with diabase seems to rule out the possibility that they represent separate intrusions. The chemical and mineral data which have been presented show a gradation (Figs. 3, 4, 7, 8, 9) between the acidic and basic facies, and this suggests a genetic relationship. It is believed that the theory of separate intrusions may be safely dismissed for the Endion and Lester River sills. Outcrops of the more irregular Northland sill are not complete enough to indicate its relations.

RED ROCK BY METAMORPHISM OF RHYOLITE

Although exposures are practically continuous, the exact contact of the intrusive red rock and the extrusive rhyolite is difficult to determine at the top of both the Endion and Lester River sills. There obviously

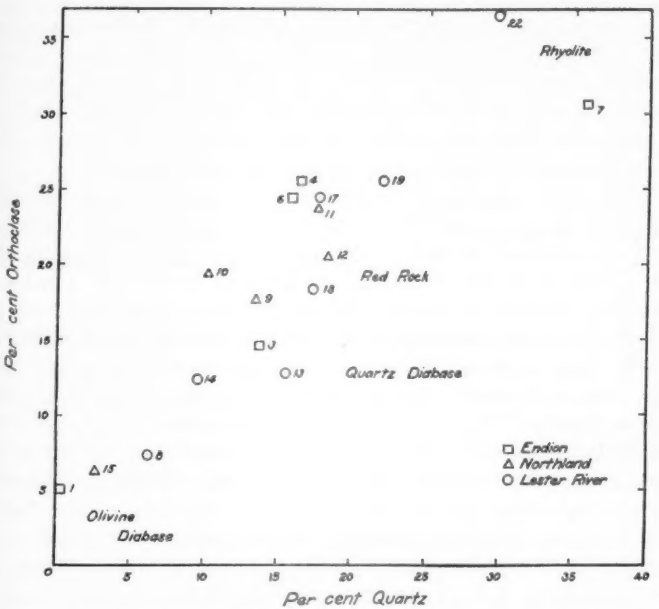


FIGURE 9.—*Quartz and orthoclase content of the norms of the Duluth sills*

has been recrystallization of the rhyolite at the contact so that it is almost as coarse-grained as the granophyr of the sills. There is, however, a sharp break in composition between known red rock and known rhyolite. This is brought out by the analyses of Table 1, by the petrographic descriptions, and by the various diagrams. (*See especially* Figures 2, 3, 4, and 7.) Thus while it was not everywhere possible to locate the exact contact between granophyr and recrystallized rhyolite, it is no doubt possible to do so by close study of doubtful specimens. Rhyolite is very high in quartz and feldspar, low in ferromagnesian minerals, and especially low in magnetite and ilmenite. Granophyr on the other hand has more ferromagnesian minerals and is notably high in magnetite and ilmenite. It may therefore be concluded that a small

amount of granitic rock at the upper contact of the sills represents metamorphosed rhyolite but that the main red rock of the sills could scarcely have formed in that way. The break in chemical composition between known granite and rhyolite is much too abrupt.

RED ROCK BY HYDROTHERMAL ALTERATION

It has been argued recently by Bastin (1935; 1938) that certain aplites and so-called intermediate rocks associated with diabase have formed by hydrothermal alteration. It must be made clear, however, that the aplites or red rocks of the Cobalt district referred to hydrothermal alteration by Bastin (1935) are different in form and structure, if not in composition, from the extensive red rock masses of the Duluth sills.

There is clear evidence of hydrothermal alteration of the Duluth sills but it is by no means confined to the red rock or intermediate rock but is fully as important in much of the diabase as well as in the associated flows. Furthermore the alterations do not tend to produce a red rock except that oxidation of iron to hematite at places produced a very red rock (No. 19) along fractures in an otherwise greenish acidic rock (No. 13). In these examples the fundamental nature of the rock is little changed except in color. Bastin (1935) emphasized albitization in the Nipissing diabase. There is no evidence of extensive albitization in the Duluth sills, and one has only to observe the relatively constant amount of soda (Figs. 4 and 7) in all facies of the three sills to realize that albitization is unimportant. Comparison of the detailed data on the Duluth sills with data on the Nipissing diabase shows other significant differences.

Comparison of the data with the changes listed by Bastin (1938) for the Pigeon Point sill also fails to support the hydrothermal hypothesis for the Duluth sills. The introduction of water, of carbonates, and of albite are emphasized by Bastin. Again study of the analyses of Table 1 shows that hydration is no more important in the red rock facies than in the diabases. Carbonation is unimportant as is shown by the relatively small amounts of carbon dioxide even in thin sections. The lesser amount of lime in all facies other than the diabase indicates a lack of carbonation, and a lack of important albitization in the Duluth sills is shown by the normal soda content of the analyses. Unfortunately chemical analyses of the Pigeon Point and Nipissing specimens described by Bastin are not available for comparison. Available analyses of the Pigeon Point rocks (Grout, 1928, p. 560) show that the intermediate rock has a modest content of soda and low lime as compared with the gabbro. It may be concluded on the basis of available evidence that none of the facies of the Duluth sills owe their main characteristics to hydrothermal

alteration, and the writers doubt that the intermediate rock at Pigeon Point was so formed.

It is desirable in this connection to point out that the expression "hydrothermal alteration" is now used by some geologists in a very inclusive sense. The hydrothermal alteration described for the Duluth sills is a result of solutions which affected the rock after consolidation, possibly long after. Deuteric effects may have played some part in the formation of the red rock. This reminds one of some of Fenner's (1931) discussions. Therefore, in concluding that the various facies of red rock in the Duluth sills did not originate by hydrothermal solutions, the writers do not even suggest that there were no hydrothermal and deuteric effects, but there is no evidence that these processes were dominant in forming the very thick red rock zones in the sills. If hydrothermal alteration were responsible, some of the basalt flows would have been more susceptible to alteration to red rock than the massive diabase.

RED ROCK BY SYNTEXIS

Daly (1917) argued strongly for the origin of the granite of the Pigeon Point sill by syntectic differentiation. The fact that the Duluth sills are closely associated with rhyolite as well as basalt flows and that the roof of both the Endion and Lester River sills is rhyolite suggests that these diabases may have derived their red rock tops by assimilation and later differentiation—that is, by syntectic differentiation as proposed by Daly.

It is difficult to obtain positive evidence for or against the theory of syntectic origin of the red rock facies of these sills, as there are very few inclusions of recognizable rhyolite in them. In other words, if assimilation was active it rather completely assimilated all fragments of rhyolite. Other basic intrusives in Minnesota do show partly recrystallized rhyolite inclusions (Ranta, 1939). Mineralogical and chemical studies as shown by the descriptions, by Table 1, and by various figures prove that there is a sharp chemical and mineralogical break between the most acidic red rock of the sills and the rhyolite of the flows. This sharp break, however, does not in itself disprove syntexis. To do so requires detailed comparison of the analyses to see whether or not some possible combination of diabase and rhyolite may not account for the composition of the red rock (granite) facies. A study of Table 1 readily shows no significant consistent differences between the amounts of alumina and soda in the sills and the amounts in the rhyolite. Calculations for iron oxide, silica, and potash show that a combination of equal parts of diabase and rhyolite would give values approximately those shown in some of the red rock facies.

While this does not in any sense prove syntexis it does admit of its possibility insofar as the constituents mentioned are concerned.

Magnesia, lime, and titanium oxide present a different problem. In the Endion sill magnesia forms 6.52 per cent of the diabase and from 2.00 to 3.74 per cent of the red rock facies, but the rhyolite above contains only a trace. This extreme is approached by the Lester River sill. To produce a red rock with 2 per cent or less of magnesia would therefore require that a given amount of diabase assimilate from equal amounts to twice as much rhyolite. Lime and titanium furnish parallel examples. Similarly potash, which is abundant in the red rock and rhyolite, is low in diabase. The diabase of the Endion sill would require the addition of 1½ times its own weight of rhyolite to furnish the potash found in the red rock facies.

If it were granted that such assimilation took place then minor constituents such as phosphorus would present difficulties. In the Lester River sill this is especially true. The amount of P_2O_5 in three facies of the diabase averages 0.54 per cent. Three facies of red rock contain 0.50 per cent, but the rhyolite above contains only .03 per cent. Under the syntexis theory the P_2O_5 of the red rock should not exceed 0.27 per cent. Likewise, MnO is very low in the rhyolite but is present in much greater amounts in the diabase and red rock.

Mineralogically the improbability of syntexis is suggested by the fact that magnetite-ilmenite appears to be about as abundant in the red rock facies as in diabase, whereas it is insignificant in rhyolite.

It may be concluded therefore that, while syntexis cannot be excluded as a process in the Duluth sills, it appears improbable that it was the dominant process. The chemical irregularities are too great.

In explaining the Pigeon Point sill Daly (1917, p. 43) emphasized syntectic differentiation rather than pure syntexis. In the Duluth sills there seems to be no direct trace of material incorporated into the magma. If any was incorporated it has changed so much during differentiation as to be unrecognizable. The differentiation is clear, but there is an entire lack of evidence of any earlier processes.

RED ROCK BY DIFFERENTIATION

The origin of different facies within intrusives by differentiation has been widely accepted, but so many alternatives have been suggested that a critical examination of the facts for each intrusive is demanded. Probably the most significant fact regarding the Duluth sills is that they display essentially only two rock facies—diabase and granite—in spite of rather uniform chemical gradation from base to top (Fig. 10). All variations within the primary rocks of the sills are varying mixtures of these

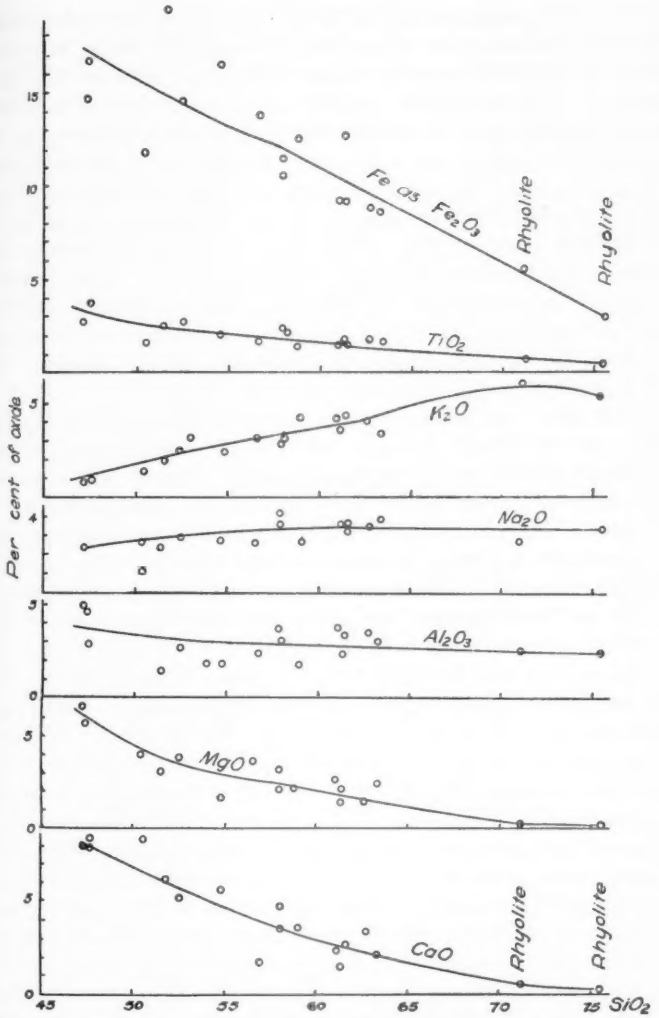


FIGURE 10.—*Variation diagrams of the Duluth sills and overlying rhyolite flows*

fundamental rock types. A somewhat similar situation in a sill in Ontario has been described by Emmons (1927). Of the 50 thin sections examined not one closely approached a diorite or syenite. The fact that

quartz or micropegmatite appears in the diabase as the first evidence of gradation to the acidic facies shows how unlikely would be the appearance of diorite and syenite. Grout (1918) has emphasized the same condition in the Duluth gabbro lopolith and suggested immiscible separation as a possible mode of origin. The lack of intermediate rocks has been repeatedly pointed out, and recently Fenner (1937) has commented on its significance. Whatever the mode of origin, this separation into gabbro and granite is characteristic of the Keweenawan rocks of the Lake Superior district. The writers have examined hundreds of thin sections of Keweenawan flows from the north shore of Lake Superior, and these without exception fall either into basalt or rhyolite classifications. Throughout the great thickness of Keweenawan flows of the north shore there are rhyolites irregularly distributed with the basalts. Such rocks as trachyte and phonolite are unknown.

This situation may represent a line of descent referred to by Kennedy (1933), but the Duluth diabases have enough olivine to suggest that they should belong to the quartz-free series, whereas the acidic differentiates are characterized by quartz and alkali feldspar (Fig. 11).

It is not probable that differentiation of the sills was a simple process. The heterogeneous mixture of acidic and basic material in the upper half of the sills indicates that, whatever the process of segregation, separation of acidic and basic material into heavy and light layers was imperfect. Interstitial micropegmatite in the lower portion of the Lester River diabase indicates an early tendency toward a more acid liquid residue. That basic and acidic portions of the sills are partly arranged according to specific gravity is clear from the rock types. Specific gravity determinations of many specimens show a range from 2.979 for the diabase near the base of the Endion sill to 2.697 for the red rock of the same sill, a variation of approximately 10 per cent. This gravity stratification may be an outcome of such processes as crystal settling, liquid immiscibility, or the concentration of mineralizers in the upper portion of the sill. The presence of a slightly lighter diabase in the lower portion of the Lester River sill may be due to chilling of the primary magma followed by segregation for the main portion of the sill.

It is noteworthy that none of the analyzed red rocks of the sills contain as much silica as the average red rock of the gabbro (Grout, 1918, p. 642) and that none of the facies of the diabase of the sills show the low silica of the ultrabasic facies of the gabbro. From these comparisons it may be concluded that the sills are not late offshoots of the gabbro but rather intrusions of an undifferentiated magma probably like that which filled the Duluth gabbro chamber. The enormous gabbro required an extremely

long period of cooling with the consequent differentiation to more extreme facies than occur in sills (Grout, 1918); the sills, being of modest size, cooled in a shorter time, and differentiation was permitted only to a modest degree.

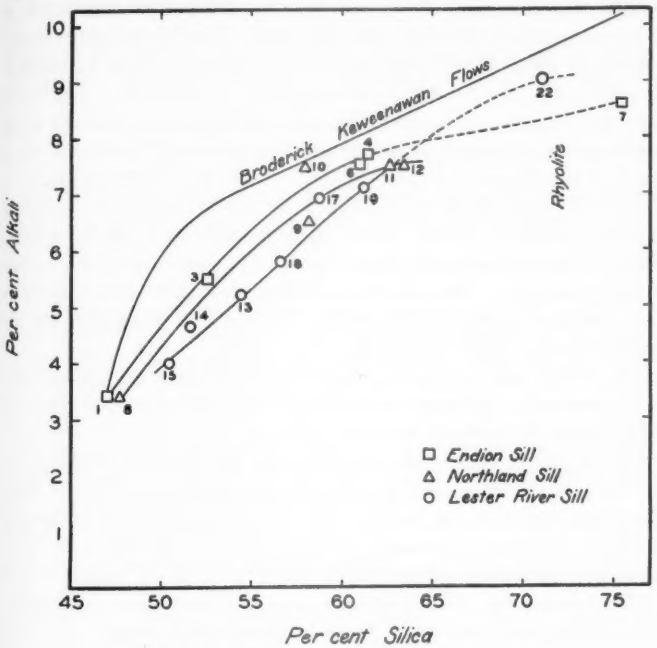


FIGURE 11.—*Alkali-silica ratios of the Duluth sills*
With curve of Broderick's Keweenawan flows for comparison.

The objections to crystal settling as the principal method of differentiation of the Duluth gabbro and the Duluth sills have been stated by Grout (1918). Difficulties with the theory of crystal settling have also been frequently pointed out by Fenner (1926; 1937). These objections are valid insofar as the present study is concerned.

Differentiation of the sills into diabase and granite facies is probably closely related to the concentration of mineralizers in the upper portion of the sills; the mineralizers played a considerable role especially in the formation of the red rock with micropegmatite as an important constituent and possibly also in the crystallization of considerable magnetite-ilmenite in the acidic facies. The significance of water was noted by Grout (1918, p. 657) in his original argument for immiscible separation.

Fenner has been a persistent advocate of the idea that mineralizers play an important role in igneous rocks and may even perhaps be responsible for the formation of some types by hydrothermal metamorphism or, what seems more likely, deuteric action (Fenner, 1926, p. 753). In the Duluth sills it seems clear that, however the granophyr originated, it was by a process which was essentially igneous. There is a lack of structural conditions which would have restricted the action of later, extraneous hydrothermal solutions to the upper portions of the sills. The rocks above are, if anything, more pervious. Although hydrothermal alteration has been somewhat active throughout the sills, it has not tended to produce a red rock in the diabase portion. It seems that the conditions found in the Duluth sills may be better explained by considering that a concentration of mineralizers, mainly water, carried acidic and alkalic constituents into the upper portion of the sills and there assisted in the formation of the granophyr. The mineralizers may well have been important in aiding the segregation of the magma previous to concentrating the material in different parts of the sill.

Emmons (1940) has published suggestions regarding the role of differential pressures in magmatic differentiation and has indicated the bearing of these ideas on the red rock problem.

The writers cannot agree with the statement by Emmons (1940, p. 11) that the sills at Duluth show less differentiation than the Pigeon Point sill. Detailed maps of the Duluth sills when compared with those of Pigeon Point (Grout and Schwartz, 1933) show that the red rock forms fully as large a percentage of the Endion and Lester River sills as it does of the Pigeon Point sill. If fracturing aids segregation, there is no reason why the Duluth sills should not have large red rock zones. All three sills intrude rhyolite which forms the roof of the Endion and Lester River sills. Along the north shore of Lake Superior rhyolite is always conspicuously jointed. There are, however, sills on the north shore (Grout and Schwartz, 1939, pls. 1-5) which have little associated red rock, and possibly further work might show that they satisfy the conditions postulated by Emmons.

The mechanics suggested may have had some effect in the remarkably good segregation of acidic material described in this paper, but direct evidence would seem difficult to get.

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GLACIATION IN THE WASATCH PLATEAU, UTAH

BY EDMUND M. SPIEKER AND MARLAND P. BILLINGS

CONTENTS

Page
Abstract ... 1173
Introduction ... 1174
General geology ... 1175
Physical character and setting ... 1175
Stratigraphy ... 1175
Structural and physiographic setting ... 1177
Joe's Valley graben ... 1177
Gooseberry-Flat Canyon area ... 1179
Glaciation ... 1181
General nature of evidence ... 1181
Cirques ... 1181
Widened valleys ... 1183
Moraines ... 1183
General character ... 1183
Compound moraines ... 1184
Simple terminal moraines ... 1187
Recessional moraines in cirques ... 1187
Glacial features south of the principal area ... 1188
Distribution of glaciers ... 1190
Relations between glaciation and faulting ... 1192
Date of glaciation ... 1193
Evidence in Wasatch Plateau ... 1193
Comparison with other glaciated regions ... 1193
Works to which reference is made ... 1196

ILLUSTRATIONS

Figure Page
1. Index map of Utah ... 1176
2. View looking north from central moraine ... 1185
3. East side of northern moraine near Staker Creek ... 1186

Plate Facing page
1. Features of glacial erosion ... 1182
2. Compound moraines ... 1183
3. Moraines in cirques ... 1188
4. The Horseshoe and Snow Lake ... 1189
5. Geologic map of part of the Wasatch Plateau, Utah, showing glacial features. 1198

ABSTRACT

The Wasatch Plateau, northernmost of the High Plateaus of Utah, was occupied by Pleistocene glaciers. The plateau is a high, rugged mass, deeply dissected. In the glaciated areas the bedrock formations are sandstone, shale, and limestone of

late Cretaceous and early Tertiary age. Here, in general, the rocks lie nearly flat, but they are cut by systems of normal faults among which graben are prominent. The drainage system evolved in this bedrock setting gave rise to a geomorphic foundation on which a somewhat unusual group of glacial phenomena developed. The largest group of glaciers issued eastward from deep canyons in the main western body of the plateau into a north-south graben valley, where they coalesced to form large, sheetlike compound moraines. The cirques and other erosional evidences of the glaciation are similar to those usual in the Cordilleran region excepting peculiarities of outline produced by the flat-lying, slightly resistant sedimentary rocks of the plateau, as contrasted with the usual mountain assemblage of strongly disturbed hard rocks. The glaciers were localized, in part independently of altitude, probably in large part because of snow distribution similar to that of today. The glaciation was clearly later than most of the faulting in the plateau, but in the central and southern parts some faulting has occurred since. All moraines so far recognized are unmistakably Wisconsin in age. There is no evidence of older glaciation comparable to that in the nearby Wasatch and Uinta Mountains, where pre-Wisconsin ice was more extensive than the Wisconsin; this raises questions that are only partly answered.

INTRODUCTION

The effects of Pleistocene glaciation have long been recognized in the Wasatch and Uinta Mountains (Atwood, 1909), but little has been said about glaciation in the High Plateaus to the south of these mountains (Fig. 1), and that little has tended to minimize the importance of Pleistocene ice rather than to emphasize it.[1] Dutton, for example (1880, p. 35, 41-42, 263, 264, 270, 279, 285), mentions glacial features in the Fish Lake and Aquarius plateaus, but comments repeatedly on the general absence of glacial evidence (p. 41-42). Gilbert, climbing over the southern parapets of the Wasatch Plateau in 1875, on his return from the memorable season spent largely in the Henry Mountains, wrote in his notebook on Sept. 9: "It is possible that some of the surface I ascribe to slides is morainal, but there are surely no glacial phenomena of note." For the country traversed by Dutton and Gilbert this general conclusion is valid, but had either of these pioneers travelled across the central or northern parts of the Wasatch Plateau, their reports would have been different. Parts of the Wasatch Plateau were strongly glaciated in Pleistocene time.

In 1930, while engaged in general geologic mapping of the Wasatch Plateau, the writers came upon clear evidence of this glaciation, and mapped the moraines and other features as completely as possible under the general conditions of work. In 1931 and following years the senior author was able to supplement and extend the work, but in general a thorough study has been impossible, and this paper is not presented as a finished product—rather as an announcement of the glaciation, with description of the most important evidence and mention of certain problems that have arisen. The paper is restricted further in that the mapping

[1] Since this manuscript was written, Gould (1939) has presented an interesting description of glaciation in the Aquarius Plateau that is notably different from that described in this paper.

of glacial features was not extended beyond the district in the northern part of the plateau where the moraines are most extensive; this area is shown on Plate 5. A few features of other parts of the plateau are mentioned, but in the main the paper is centered on the area mapped.

The writers are indebted to W. C. Alden and W. H. Bradley for critical suggestions respecting the original manuscript of this paper.

GENERAL GEOLOGY

PHYSICAL CHARACTER AND SETTING

The Wasatch Plateau, northernmost of the High Plateaus of Utah, lies in the central part of the State (Fig. 1). It is a true table land, deeply dissected by streams tributary to the Colorado River on the east and to Sevier River and Utah Lake on the west. Remnants of the flat crest, most of which are long and narrow, stand between 10,000 and 11,000 feet above sea level in all except the northern part of the plateau, where the upland surface is nearer to 9000 feet. Sanpete Valley, on the west, and Castle Valley, on the east, range in the main between 5000 and 6000 feet in altitude, and the average relief is thus about 5500 feet. The maximum relief is about 6200 feet.

Throughout the plateau, excepting the western and northern margins, the strata lie nearly flat or dip at moderate angles, and resistant rocks alternate with less resistant; the land forms carved out of this mass are therefore characterized by the cliff, bench, and slope profiles common in the Colorado Plateaus. The upper levels of the plateau are well watered, and the profiles there are somewhat rounded, but the borders are typically semi-arid in climate, and their profiles are angular.

The drainage pattern of the plateau is in general dendritic, except where affected by faulting as described farther on. The main divide, which is part of the major divide between the basin of the Colorado River and the Great Basin, is near the western border in the northern part of the plateau, but in the southern part it swings to a more nearly central position. The east-flowing streams in the northern and central parts are therefore much longer than the west-flowing. On the east are the headwaters of Price, San Rafael, and Muddy rivers, the major tributaries of the Colorado in this part of Utah. On the west are short tributaries of San Pitch River, and longer tributaries of Sevier River; these streams join to flow into Sevier Lake. In the plateau all these streams flow in deep canyons, and the surface features are characterized by bold relief.

STRATIGRAPHY

The rocks of the Wasatch Plateau have been described in several publications (Spieker and Reeside, 1925; Spieker, 1931; Spieker and Baker,

1928) and for present purposes a bare outline of the stratigraphic column will suffice, with comments on those rock characters that are important in glacial phenomena. It should be noted that part of the section has re-

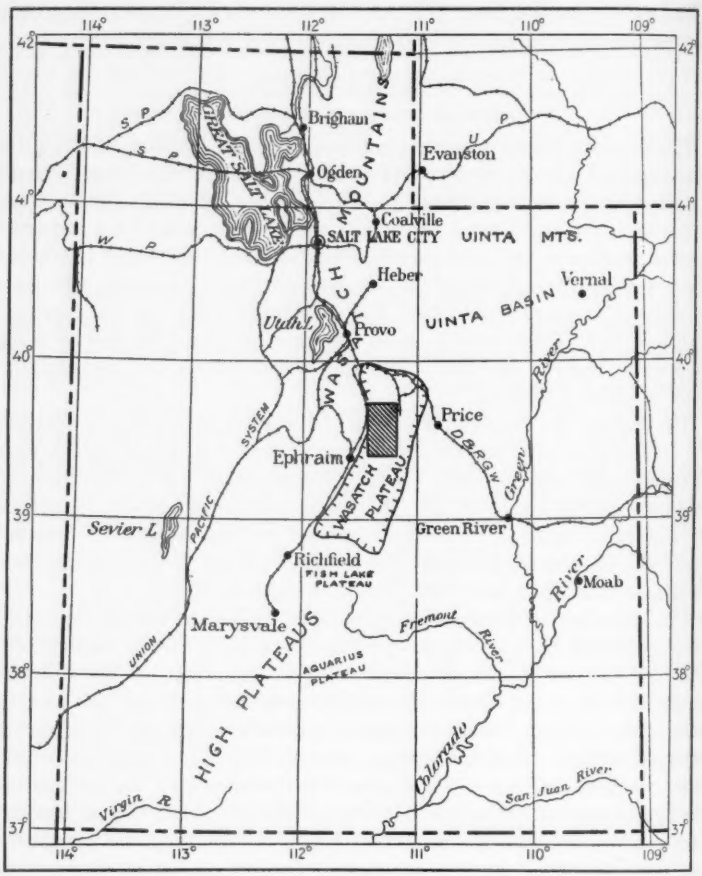

FIGURE 1.—*Index map of Utah*

Showing location of Wasatch Plateau. Shaded rectangle is area covered by Plate 5.

cently been revised (Spieker, in preparation). The formations present in the part of the plateau shown on Plate 5 and here especially described are outlined as follows:

Stratigraphic section, north central part of the Wasatch Plateau

Paleocene (?)	
Flagstaff limestone. Gray, tan, white limestone, with minor amounts of shale and sandstone; lacustrine	300–500
Upper Cretaceous and Paleocene	
North Horn formation. Buff, gray, red sandstone, gray to variegated shale, conglomerate, some limestone; flood-plain and lacustrine in origin	2000
Upper Cretaceous	
Price River formation.	
Upper member: Gray sandstone and conglomerate with minor amounts of shale	600+
Castlegate sandstone member: Massive, cliff-forming gray sandstone, coarse-grained to conglomeratic	300
Blackhawk formation. Medium-to fine-grained buff and gray sandstone, gray shale, coal	1500
Star Point sandstone. Massive cliff-forming buff sandstone, medium-to fine-grained; marine	450
Mancos shale. Gray marine shale (only uppermost part exposed in area here described)	4000+

The thicknesses given apply only to the part of the plateau described in this paper. The Mancos, Star Point, Blackhawk, and Price River formations and the lower two-thirds of the North Horn formation are Upper Cretaceous in age; the upper part of the North Horn is Paleocene, and the Flagstaff limestone may also be.

In the glaciated parts of the plateau the North Horn formation is the dominant bedrock at the surface. The shale and sandstone of this formation are not very resistant, a fact of some importance in the interpretation of erosional features and moraines produced by the glaciers. In the part of the plateau especially described in this paper the cirques have all been carved out of North Horn strata, but in areas to the north and south the heads of glaciated valleys are in the Flagstaff limestone, whose greater resistance to physical destruction has yielded somewhat more angular profiles, modified on the other hand by the characteristic chemical weathering of limestone. The Price River formation, in which many of the glaciated valleys have been cut, consists of sandstone more resistant than either of the two overlying formations; this has given rise to steeper valley sides and more abundant cliffs and ledges.

STRUCTURAL AND PHYSIOGRAPHIC SETTING

Joe's Valley graben.—The setting in which the glaciers developed was determined by a combination of structural and physiographic features best described as a unit.

The gently warped strata of the plateau are broken by groups of normal faults, of which the more important commonly form graben. The longest and most prominent of these sunken blocks is the Joe's Valley graben, which begins in the north central part of the plateau, and extends (nor-

mally 2 to 2½ miles wide) more than 60 miles in a general direction slightly east of south through the central and southeastern parts of the plateau. The northern part of the graben is shown on plate 5. Throughout its length the dropped block is complicated by minor faults, and the major marginal faults are offset in places to form en echelon patterns; one such occurrence is evident on the western margin of the part of the graben shown on Plate 5. The maximum displacement ranges between 1500 and 3000 feet, and in the northern part it is 1400 to 1600 feet on the average. Physiographically the graben is strikingly expressed throughout its course in the plateau; it is a well marked trench, abruptly walled in by the scarps of the opposing faults. In the part of the plateau here described the graben block is generally tripartite, with a central ridge bordered by valleys on either side that parallel the bounding scarps. The central ridge is locally a horst, elsewhere purely erosional in origin; the flanking valleys follow the zone of shattered and weakened rocks adjacent to the major faults. Some of these valleys are broad and flat, as at Upper Joe's Valley, Sheep Flats, and Miller Flat.

The drainage system of the graben gives evidence of an interesting erosional history. The longitudinal streams in the graben head against transverse divides from which they flow alternately northward and southward to find issue eastward through pronounced canyons in the high block east of the graben. There is strong contrast between the streams flowing into the graben from the high lands to the east and those from the west. The eastern streams are short and ephemeral, hardly a mile long, whereas those on the west side are 3 to 4 miles long, and are all perennial. The western valleys were glaciated, the eastern were not. This drainage development is the result of interplay between faulting and erosion whose major features must be clearly in mind if the significance of the glacial evidence is to be understood.

The drainage on the eastern slope of the plateau is normally dendritic, tributaries converging toward major streams which flow eastward and southeastward out of the plateau to join the Colorado. The major streams flow directly across the Joe's Valley graben, showing little sign of its influence; they are probably antecedent. The drainage pattern has been locally affected by the faulting, however, and along the graben it has been partly dismembered, tributary systems to the west having been diverted to follow the downfaulted trench. The northern headwaters of the Left Fork of Huntington Creek, for example, converge out of a group of canyons—Spring, Lake, Rolfson, and Staker—(Pl. 5), that head against the crest of the plateau, and flow into the graben trench. Here they unite and cross the fault zone with little adjustment, plunging out of the trench into a rocky gorge about 3000 feet deep, which the Left Fork follows to

Huntington Creek, not far east of the area shown on Plate 5. The next group of canyons to the south, however—Jordan, Seely, and Bennett canyons—formerly drained eastward directly across the site of the graben through the present course of Horse Canyon, to join Huntington Creek not far east of the margin of Plate 5. The notch of this old drainage channel is plainly evident in the eastern scarp of the graben, and may be seen in the configuration of the outcrops on Plate 5. This system was dismembered by the faulting; the headwaters were diverted northward into the trench to join the Left Fork. Likewise the next group of streams —Potter's, Baconrind, Brown's, and Bulger creeks—which head against the main divide, formerly united to form a creek that flowed southeastward across the graben in a channel even more clearly evident than the Horse Creek gap that lies just south of the boundary of Plate 5. This drainage formed the northern headwaters of Cottonwood Creek, the next main stream south of Huntington Creek (See Hiawatha topographic sheet, United States Geological Survey, or Pl. 31, Spieker, 1931) and was diverted southward into the trench, to join Seely Creek, the central tributary of Cottonwood, where it crosses the graben.

The surface features thus evolved by interaction of faulting and erosion, and especially the group of ten canyons in the western table of the plateau debouching into the trench of the graben, formed the setting in which the glaciers developed and built the assemblage of compound moraines described in the following parts of this paper.

Gooseberry-Flat Canyon area.—North of the area just described the main crest of the plateau drops off from altitudes around 10,000 feet to levels near 9000 feet, and the configuration of the plateau changes. The Joe's Valley graben ends at Cleveland Reservoir, against a range of hills which is the northwestward continuation of the mountains east of the graben. The abutment of these hills against the trench is caused mainly by a fault which trends northwestward from the eastern Joe's Valley fault, against which the western faults all terminate, stopping off the graben. This fault continues northwestward through mountainous territory where it is not plainly expressed in surface features until it reaches the head of Flat Canyon. There it is marked by a scarp which curves into the north-trending scarp of the Gooseberry fault, which forms the prominent eastern wall of Gooseberry Valley, a broad, open, alluvium-filled depression that lies between the narrow western divide of the plateau and the somewhat higher and broader table land north of Flat Canyon. This valley, together with Flat Canyon and the upper course of Gooseberry Creek, contains evidence of an erosional history whose relation to the faulting is somewhat different from that of the Joe's Valley graben.

The facts involved are possibly important in the problem of older glaciation. Although this problem is specifically brought up later in the paper, it is perhaps well to summarize the general physiographic evidence here, that it may conveniently be related to the history of the Joe's Valley graben.

Gooseberry Creek heads in a small cirque northwest of Spring Canyon and flows northward through a broad valley, whose gently curving bottom is interrupted only by the Gooseberry moraine, until it reaches lat. 39° 40′, where it plunges into a small, tortuous gorge in a motley group of rounded hills. Southeast of the Mammoth Ranger Station it issues from this gorge into the broad valley mentioned above, where it meanders northward for about 2 miles and then turns abruptly westward into another gorge which swings around to a northeastward trend and becomes a deep canyon where it crosses the Gooseberry fault.

Flat Canyon is strikingly abnormal. As its name implies, its bottom is flat, its cross profile broadly U-shaped, and these outlines continue upstream to its very head, where the U-profile is transected by the scarp of the cross fault described above. At this scarp the surface descends from the valley bottom abruptly into the headwaters of a tributary of Gooseberry Creek. The valley of this tributary, also fairly broad, widens out into the open spaces of the upper Gooseberry valley. Although surface expression of the cross fault is not marked elsewhere, it seems possible that here the valley bottom has been displaced. Alternative explanations, which need not be discussed here, introduce physiographic complexity and seem in some ways less probable.

With these facts in mind, a brief perusal of the map (Pl. 5) will suffice to bring out a clear picture of drainage diversion. The present upper part of Gooseberry Creek formerly flowed out through Flat Canyon and was separated from the part now downstream by a divide in the low hills above described. The next section of Gooseberry Creek formerly constituted the headwaters of Huntington Creek, and flowed out southeastward through Brooks Canyon. Barbed tributaries, involving details too small to show on the map, afford convincing evidence of this, although the picture is not entirely simple and some facts are difficult to reconcile. Brooks Canyon is a rocky gorge, and the opposing head of Huntington Canyon is likewise narrow, but where it turns into its dominant southeasterly course Huntington Creek enters a flat-bottomed valley which continues open and fairly wide as far as the word Huntington on Plate 5; there the valley narrows and becomes a deep mountain gorge.

This unusual combination of drainage features bears on two aspects of the present subject: First, it provided the setting in which the late Pleistocene glaciers developed, and second, the combination of broad upland val-

leys, U-shaped canyons, and lower V-shaped gorges is not easy to account for by any hypothesis, but it suggests older glaciation, as discussed farther on.

GLACIATION

GENERAL NATURE OF EVIDENCE

The normal features produced by mountain glaciation, namely, cirques, U-valleys, and moraines, are present in all parts of the high western crest of the Wasatch Plateau, but in the southern half of the plateau the cirques might be interpreted as amphitheatres produced by weathering and erosion after the fashion discussed by Hinds (1925) and Freeman (1925), and the morainic material as landslide debris, the interpretation favored by Gilbert in the quotation given at the beginning of this paper. Only in the northern half of the plateau are the phenomena, and especially the moraines, prominent enough to leave no doubt as to their mode of origin. In the area covered by Plate 5 all the usual results of mountain glaciation are clearly displayed.

Although the phenomena are perfectly normal, viewed in terms of ordinary glacial action, they are different in some respects from the usual picture presented by glaciated areas in the North American Cordillera. The combination of bedrock geology and geomorphic development described in foregoing paragraphs has given rise to cirques that are commonly broad and shallow, in some cases almost angular in plan, instead of deep and bowl-shaped; the widened valleys below the cirques are broader and flatter in cross-profile than the usual glaciated canyons, and the principal moraines, pushed out into the longitudinal graben trench, are compound and sheet-like, rather than crescentic or looped; lateral moraines are inconspicuous or absent. Finally, the small recessional moraines in the cirques are more clearly developed than in any other region known to the authors, possibly excepting Glacier National Park (Alden, 1914).

In the following descriptions the attempt is made to give just enough detail to characterize the evidence adequately for the uses of geologists interested in the general problems of Cordilleran glaciation. Fuller description, even where possible, would be incommensurate with the extent to which field studies have been carried.

CIRQUES

Between latitudes 39° 25′ and 39° 40′, a distance of about 14 miles, the crest of the plateau is scalloped on the east side by a succession of glacial amphitheatres, 17 in all, various in size and form but consistent in prominence. The general pattern produced by these cirques stands in contrast to that on the west side of the divide, where the canyons are

generally V-shaped. This contrast is plainly evident in the outcrop pattern on Plate 5. In the plateau as a whole there are a few cirques in north-facing heads of streams on the west side, as for example the one not far south of lat. 39° 30′, but in general the absence of glacial features on the west slope is worthy of note. The distribution of the glaciers in general is discussed later in this paper.

The cirques vary considerably in outline, as is shown by the hachuring on Plate 5. A few are semicircular amphitheatres with rounded bottoms, as for example the two heads of Seely Canyon, but most of them are broad, shallow, and flat-bottomed; some, such as the heads of Boulger and Spring Canyons, are even somewhat angular in outline. Etched out of a nearly flat plateau surface, the cirques have fairly sharp rims, with very steep slopes plunging down to broad, flat bottoms. They are in general not overdeepened, and the small lakes present in many of them (not all mapped) are the result of morainic damming rather than gouging. Some of the cirques, such as Jordan, Rolfson, and the north fork of Lake Canyon, are compound, the divides between two or more head tributaries having been worn away by the ice. The east-west divides between canyons are in general narrow and flat-topped or rounded, but some are sharp and at a few places cols have been cut by the mutual approach of two névé fields; examples of this are found between Rolfson and Staker canyons, and between Jordan and Seely canyons. Most of the cirques are 400 to 600 feet deep, but some are shallower, for example the groups of smaller ones at the heads of Potter's and Baconrind canyons. One of the more symmetrical shallow cirques, the head of the South Fork of Lake Canyon, is shown on Plate 1, Fig. 1.

The cirque floors are generally covered with morainic and alluvial debris, and are not rocky as are the cirques in mountains formed of igneous, metamorphic, or other hard rocks like the Sierra Nevada and the Uinta Mountains. Direct evidence of scouring is thus largely lacking. The cirque floors, and in part the valley floors below are characterized by steps, some of which are more than 100 feet high at the shoulder. This phenomenon is common in glaciated mountains the world over (Hobbs, 1911, p. 59-69; Russell, 1891; Johnson, 1904); it seems generally to be due to glacial quarrying by the process known as stoping, a combination and sequence of plucking and sapping of the type active in the bergschrund. The crevasses developed at ice-falls afford occasion at such places for the intense weathering that apparently plays a major role in the sculpture of glaciated depressions (Johnson, 1904). The production of these steps was obviously favored by the nearly flat-lying attitude of the sedimentary rocks in the Wasatch Plateau. The most prominent and extensive steps are in the southern part of the plateau, where the glacial

FIGURE 1. CIRQUE AT HEAD OF SOUTH FORK OF LAKE CANYON
A typical shallow, flat-bottomed amphitheatre, about 400 feet deep. Moraine rampart at extreme right.

FIGURE 2. SPRING CANYON AT HUNTINGTON RESERVOIR
An abnormally widened valley. Looking southeast; Joe's Valley graben in distance.

FEATURES OF GLACIAL EROSION

FIGURE 1. MORAINE TOPOGRAPHY NEAR CLEVELAND RESERVOIR
Looking southwestward across northern end of Joe's Valley graben. Moraine in foreground; bedrock hill beyond.

FIGURE 2. NORTHERN LOBE OF CENTRAL MORAINE AT MILLER FLAT
Looking southeast. Hill to left is bedrock; moraine is accentuated by dense coniferous forest.

COMPOUND MORAINES

features were not mapped, but here, where the evidence in general is not so decisive, it is not known to what extent the steps are due to ice action alone; they may be due in part to the other agents of denudation that have produced the normal step-like profiles of the plateau.

WIDENED VALLEYS

In the principal glaciated area shown on Plate 5 the canyons below the cirques have been widened to the U-shaped profile characteristic of glaciated valleys. Many of these valleys are of normal proportions, but some, such as the left fork of Spring Canyon (Pl. 1, Fig. 2), are abnormally broad and flat-bottomed. None of the valleys is notably overdeepened.

The unusually broad valleys are apparently due to greater ease of lateral than downward cutting by the large volumes of ice issuing from the larger catchment basins. A glacier of given volume that started to cut laterally would thereby grow thinner and lose the concentrated weight necessary for vigorous downcutting. Further, the tendency toward extensive lateral cutting is explainable in part by the greater vulnerability of the sandstones and shales concerned along the bedding than across it.

Other than the phenomenally broad valleys, the largest of which are indeed striking, the glaciated stream courses present no features worthy of special mention. Most of them are thinly veneered by morainic material, and are only slightly notched by post-glacial streams.

MORAINES

General character.—The bodies of morainic material observed in the plateau all show, small as well as large, the topographic features characteristic of glacial deposits. The surfaces are hummocky and irregular, and are dotted with hundreds of small depressions, most of which are occupied by ponds or marshes except in very dry seasons. Examples of this typical knob-and-kettle topography are shown in Plate 2. Post-glacial erosion has hardly begun to modify these features; the only noticeable effect of recent stream-work is notching on the main stream courses, and even these channels are for the most part notably influenced in course by the configuration of the moraines. This feature is plainly evident in the drainage pattern of the two southernmost moraines shown on Plate 5.

The morainic material itself is typical till, composed of ice-shaped fragments of sandstone and shale, locally also limestone, embedded in a matrix of unstratified sand and clay. The boulders in general are not large; not many exceed 2 feet in diameter. Striated and facetted boulders occur, but they are rare; the rock materials are for the most part friable, and have commonly been broken up, rather than worn down by attrition. The till

is uniformly fresh; neither the individual rock fragments nor the matrix shows notable effects of weathering.

The moraines in the area shown on Plate 5 are of four types: (1) Simple terminal; (2) compound terminal, piedmont type; (3) ground moraine; and (4) recessional moraines. No well defined lateral moraines have been observed in this area, but to the south, in Black and Reeder canyons, lateral moraines have been observed but not mapped; they are described farther on. In the area of Plate 5 the terminal and recessional moraines were mapped, but no attempt was made to outline the ground moraine in the glaciated valleys.

The distribution of these moraines is shown on Plate 5. The compound moraines are restricted to the Joe's Valley graben, and terminal moraines of the usual type are found to the north; the most prominent of these are the Gooseberry moraine and the Boulger moraine, in Flat Canyon. The recessional moraines are found almost exclusively in the cirques.

Compound moraines.—By far the most prominent moraines in the Wasatch Plateau are the three great masses of till in the western part of the Joe's Valley graben between Mill Canyon and the Huntington Reservoir. These moraines, formed by the coalescence of ice-tongues issuing into the graben trench from three groups of canyons, are essentially of the piedmont type, and are somewhat unusual among late Pleistocene glacial deposits of the Cordilleran region. They have resulted from the development of powerful glaciers in the peculiar physiographic setting whose origin is outlined on foregoing pages.

Instead of the crescentic or looped heaps of till that characterize the terminal deposits of the usual mountain glacier, the ice-tongues of this part of the Wasatch Plateau spread out extensive sheets of debris that resemble more, in general view, the moraines of the large ice-sheets than those of valley glaciers. Although locally somewhat thicker near the mouths of the canyons than elsewhere, they bear no striking evidence of concentrated accumulation at the ends of the individual ice tongues. Locally, recessional ramparts are evident, but they are low and not at all striking. The morainic masses are essentially sheets of till, 200 to 300 feet thick, characterized by knob-and-kettle topography, and as a rule by steep, well-defined frontal slopes. They clearly indicate coalescence of the contributing glaciers into ice sheets of approximately the same extent north to south as the moraines themselves.

The three piedmont moraines thus formed may be conveniently designated the northern, central, and southern moraines. The northern moraine was formed by glaciers from Spring, Lake, Rolfson, and Staker canyons, the central, from Jordan, Seely and Bennett canyons, and the southern

from Booth's, Potter's, Baconrind and Brown's canyons. The northern moraine spreads fairly symmetrically from the mouths of the contributing canyons, but the central moraine extends asymmetrically some distance northward, toward the valley of the Left Fork, and the southern moraine

FIGURE 2.—*View looking north from central moraine*

Slope on left formed by sandstone of Price River formation; Miller Flat on right. Central moraine in foreground; northern moraine in right distance.

extends southward past the mouths of Bulger and Mill canyons, which contributed no till.

The area between the northern and central moraines is shown in Figure 2, and the northern margin of the central moraine is shown in Plate 2, figure 2.

The margins of the moraines are distinctly lobate wherever the ice was free to spread without obstruction, the lobes reflecting the contributions of the individual ice tongues. This phenomenon is best illustrated by the northern moraine, in which there are pronounced lobes opposite the mouths of Staker and Rolfson canyons, and a third representing the Lake-Spring glacier. Between these lobes are large V-shaped reëntrants free of drift. The reëntrant between the Rolfson and Staker lobes is shown in Figure 3. The central moraine shows this phenomenon less perfectly because the compound ice sheet abutted in its central part against a high hill in the graben, and in configuration most of the eastern margin of this moraine was determined by the fairly smooth outlines of the obstruction. The southern moraine embodies two major lobes, and minor ones, especially at the southern end, that indicate the conformation of the ice masses into some semblance of individual streams, and not perfectly uniform sheets. The outlines of the graben topography

influenced the accumulation of the southern moraine as well as the central, in that the northern lobe was constricted by a high bedrock hill to the east, whereas south of this hill the ice was free to penetrate a breach into low lands and to extend down the main western valley

FIGURE 3.—*East side of northern moraine near Staker Creek*

Hills to extreme right, Price River formation in graben block. In foreground, interlobate area between coalescent moraines of Staker and Rolfson glaciers.

of the graben, depositing a prominent hill of moraine known as Red Pine Ridge.

In details of surface configuration these moraines are very striking; the hundreds of small ponds and marshes, interspersed with equal numbers of irregular hummocks, make an impressive picture of morainic topography. On the scale adopted for mapping it would be impossible to show all the pools and depressions, the largest of which are no more than 100 yards in diameter, and their superabundance discouraged the attempt to map them even in part. A few of the more prominent ones on the northern moraine may be seen in the southwestern part of the Scofield topographic sheet of the United States Geological Survey.

The lowest points on the margins of these moraines are generally between 8400 and 8500 feet in altitude, but the southern lobe of the southern moraine extends somewhat lower, perhaps not far from 8300 feet.

Natural dams formed by the northern moraine have been utilized, by the simple expedient of plugging up the stream-cut notches, to form Huntington and Cleveland Reservoirs, which serve to store surplus water of the spring flood season for irrigation during the summer in farming districts of Castle Valley, east of the Wasatch Plateau. The dam of Huntington Reservoir is largely formed by the pile of morainic debris left at the mouth of Spring Canyon; the abnormally widened valley (Pl. 1, fig. 2) serves effectively as a storage basin. Cleveland Reservoir is impounded in the unglaciated northern end of the Joe's Valley trench

by the northern lobe of the moraine, which extended eastward across the valley, blocking it off.

Simple terminal moraines.—Many of the smaller glaciers deposited simple terminal moraines in their channels. Some of these are difficult to distinguish from landslide debris, and in the area shown on Plate 5 the less conspicuous ones were not mapped. The best examples of the type are the Gooseberry and Boulger moraines, and these because of their relation to the preëxisting topography and their possible importance in the question of older glaciation, are briefly described.

The Gooseberry moraine is a low, sprawling, crudely crescent-shaped mass spread out in the center of a broad pre-glacial valley. Behind it is a lake, Gooseberry Reservoir, artificially dammed as were Huntington and Cleveland reservoirs. It is marked, as are many of the moraines of the area, by rather dense coniferous forest, a feature which makes it easily distinguishable from the surrounding non-morainic terrane, which is not so forested. The surface of the moraine is typically irregular. The glacier which formed it headed in a small north-facing cirque east of the main divide and was not much more than 2 miles long. Some inconspicuous masses of moraine along the course of the former glacier, perhaps true lateral moraines, suggest that the ice tongue did not fill the valley; this is also indicated by the terminal moraine. The important features of the Gooseberry moraine are its situation in the broad, flat-bottomed valley, and its altitude, the lowest point on its margin being about 8750 feet above sea level, compared with 8400 feet and less for the moraines to the south.

The Boulger moraine, so called because it was formed by a glacier that occupied Boulger Canyon, is situated in the lower part of Flat Canyon. The ice sheet issuing from Boulger Canyon pushed at right angles across the broad bottom of Flat Canyon, jammed against the north wall and concentrated the moraine in a narrow heap. The deposit blocked the passage of Flat Creek, forcing it toward the north wall, where it has cut a narrow gorge into bedrock. The volume of till in this moraine seems somewhat small, considering the size of the erosion basin and compared with the other moraines of the larger glaciers. The moraine surface displays knob-and-kettle topography. The altitude at the foot, about 8700 feet, is similar to that of the Gooseberry moraine.

Recessional moraines in cirques.—One of the most interesting of the morainic developments in the region is the presence in most of the well-developed cirques of low, rampart-like ridges of till, averaging 25 to 35 feet high and extending partly or entirely across the cirque floors. Some are perfectly straight embankments, others are curved. Small ponds,

both past and present, have been impounded behind some of them. These small recessional moraines are most variedly developed in Jordan and Staker canyons, and a fine example of the straight type is found in the head of the south fork of Lake Canyon. The more prominent ones are mapped on Plate 5; the one in Lake Canyon is shown on Plate 3, figure 1, and the Jordan Canyon assemblage is partly shown on Plate 3, figure 2. In a general view, especially from the cirque rims above, these moraines are easily distinguishable because of the belts of coniferous trees they support.

These low embankments are insignificant and would command little attention were they not so uniformly present in the cirques. They apparently indicate late stands of the ice fronts, after the major recessions of the ice tongues, which seem not to have been clearly marked by any correlatable recessional halts. Some glacierets which clustered around the cirque margins evidently stood long enough to form small moraines along their lower margins. In the Wasatch Plateau they record a recognizable stage in the glacial history, but if the conclusion is correct that the larger moraines are Wisconsin in age (see discussion of age, farther on), other glaciated districts of the Cordillera seem to contain no clear-cut counterpart, possibly excepting Glacier National Park, where Alden (1914) has observed small moraines not far below the present ice fronts that are probably very late Pleistocene or Recent.

GLACIAL FEATURES SOUTH OF THE PRINCIPAL AREA

In the higher parts of the plateau south of the area shown on Plate 5 most of the larger canyons show evidence of glaciation. Although the glacial features were not mapped in this part of the plateau, observations were made of certain occurrences that should be briefly described here.

The next two large canyons tributary on the west to the Joe's Valley trench south of the border of Plate 5, Black Canyon and Reeder Canyon, contain large masses of moraine that are somewhat different in outline and situation from those of the northern area. They are more or less concealed from general view, deep in the heavily forested canyons where they merge in part with the bedrock topography. They were not recognized in 1930, when these drainage basins were being mapped by triangulation from above. In 1934 the senior author made reconnaissance trips of one day each up the canyons, but time was not available to map the moraines. The facts obtained are summarized here.

Both Black and Reeder canyons are deep, narrow gorges where they open into the Joe's Valley graben, and their mouths are sharply V-shaped notches in the western scarp of the trench. Not far within, however, both canyons show evidence of glaciation, and they are remarkably simi-

FIGURE 1. MORAINE IN SOUTH FORK OF LAKE CANYON
View from northwest rim of cirque. Moraine is marked by line of conifers.

FIGURE 2. RECESSIONAL MORAINES IN JORDAN CIRQUE
Looking southeast from rim; Joe's Valley graben in distance. Only southern part of cirque is shown.

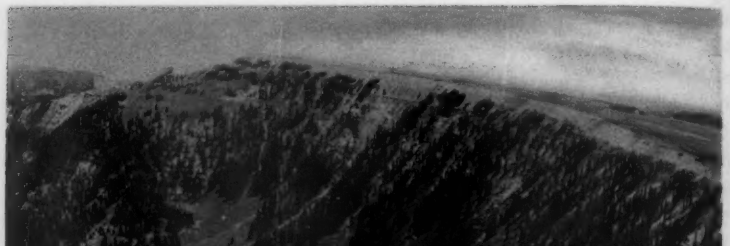

FIGURE 1. THE HORSESHOE
A pronounced, symmetrical cirque, facing north on west side of plateau east of Ephraim. Rim formed by Flagstaff limestone; altitude about 11,000 feet.

FIGURE 2. SNOW LAKE
A small cirque beheaded by fault on left that has displaced the plateau surface; scarp forms dam impounding lake. Bedrock is Flagstaff limestone.

THE HORSESHOE AND SNOW LAKE

lar in peculiarities of distribution of the moraines. In Reeder Canyon about a mile above the mouth a narrow ridge of moraine rises on the north side of the canyon and attains a height of more than 200 feet. Upstream a second, lower ridge appears next to the creek. The south wall of the canyon is formed of bedrock. Somewhat less than half a mile up the moraine ridge a sharp, steep-fronted, beautifully curved recessional (possibly terminal) moraine appears, its surface dotted with kettles and irregularly spaced hummocks. The lower ridges have the outlines of lateral moraines, but they can hardly be true laterals—there is not room south of them for any considerable body of ice, and the present V-shaped stream channel is probably the result of rapid post-glacial erosion. The till of the narrow lower moraine and that of the crescentic one above show no obvious differences in composition; they are equally fresh in appearance and there is no clear-cut evidence of great difference in age. Above these moraines Reeder Canyon is choked with glacial debris, through which the creek has cut a narrow gorge. Without this morainic material, the canyon would be wide and U-shaped, but as it is, the stream channel is narrow, V-shaped, and the lower part of the canyon, viewed from above, has the appearance of a normal, young mountain stream course, the masses of glacial debris resembling rough benches of bedrock.

In Black Canyon remarkably similar features are present. Not far within the gateway in the scarp, a longitudinal ridge of moraine appears on the north side of the canyon, and where this ridge flattens off, terrace-like, there is a recessional or terminal embankment, like the one in Reeder Canyon but not so nicely curved. Above these lower moraines Black Canyon widens out more than Reeder, and into its open spaces both lateral and terminal moraines project from the side canyons. These moraines seem almost certainly to have been deposited after the recession of the glacier in the main canyon. The lower moraine in Black Canyon might perhaps be interpreted, from its general appearance and degree of weathering, to be notably older than the ones above, but facts adequate to support a definite conclusion have not been obtained. On the north side of the canyon, about half way up, there is a lobate frontal moraine in the mouth of a tributary canyon that was almost certainly formed after the recession of the main glacier.

The moraines in Black and Reeder canyons reach lower limits similar to those of the larger moraines to the north, approximately 8300 to 8400 feet.

West of the head of Reeder Canyon, on the west slope of the plateau, is a north-facing amphitheatre known as the Horseshoe. The name is appropriate. This basin is probably the most nearly perfect in form of all the cirques in the plateau. Some suggestion of its outlines is given

by Plate 4, figure 1, but no viewpoint is available on the ground that will encompass the basin adequately in a photograph. The ice that formed the Horseshoe did not extend far below the catchment basin; the canyon below is not glaciated and the moraine at the foot of the cirque is insignificant.

South of this district glacial ice was present in all the drainage heads on the east side of the main crest, but the results of its work are not so consistently evident as in the districts to the north. A few good cirques are present, and some well-defined moraines, but on the whole both erosional and depositional traces of the glaciation are either poorly developed or in forms interpretable as the result of agents other than ice. Perhaps the most striking feature of the evidence in the southern part of the plateau is the presence of glaciated surfaces that have been offset by faulting, and the proof they afford of post-glacial movement in this part of the region.

DISTRIBUTION OF GLACIERS

The glaciers of the Wasatch Plateau were almost exclusively restricted to east- and north-facing slopes of the westernmost high table land of the plateau, the main divide.

The restriction to east- and north-facing slopes was probably due to various effects of difference in exposure. One such effect is difference in air temperatures; south-facing slopes are favored by much higher insolation values than north-facing, and the average diurnal range in air temperature, which reaches a peak just before mid-afternoon, favors the west-facing slopes as to total amount of heat received. This factor alone, however, is hardly sufficient to account for the absence of glaciers in the western canyons when long and persistent ice tongues were present just across the divide. Distribution of air currents, affecting precipitation, drifting, accumulation, and melting, must have played an important part.

The distribution of glaciers in the Wasatch and Uinta Mountains (Atwood, 1909, p. 67, 91) appears to support this conclusion. In the Wasatch Mountains the cirques are variously located, and the largest glaciers were on the west side. East- and north-facing slopes were favored in part (e. g., Big and Little Cottonwood canyons, Mt. Timpanogos) but groups of west- and south-facing catchment basins were glaciated (e. g., American Fork, Alpine Canyon). Temperature difference, the only factor common to all regions, clearly did not operate exclusively here. Nor did it in the Uinta Mountains, where the longest glaciers were on the south slope instead of the favored north slope; this Atwood (1909, p. 68) explains as due to differences in structure which allowed formation of larger catchment basins on the south side. In the Wasatch Plateau no such structural difference exists, and the catchment basins on the east and west sides

of the crest are about equal in size. The only notable difference is that the basins on the west side descend more steeply than do those on the east. The steeper slopes may have favored avalanches that prevented the accumulation of sufficient névé in the higher basins to form effective ice, but this hardly seems an important factor.

Present-day conditions in the region may throw some light on the matter. Although it is not safe to assume that in the pronounced climatic changes that have supervened since Pleistocene time all the factors influencing distribution of snow have remained constant, yet it seems that the major determinants of winter weather, dominant among which is the regional distribution of surface features, have probably not changed enough to invalidate a comparison. At all events it is true that the present distribution of snowfall and drifting of snow in the plateau is such that with intensification sufficient to produce glaciers the distribution of Pleistocene ice might easily be duplicated. Mr. J. W. Humphrey, Supervisor of the Manti National Forest, has kindly supplied available data on the distribution of snow. Although there is expectably little difference in the snowfall on the catchment basins immediately to east and west of the main divide, the strong winter winds, blowing dominantly to the north and east, cause the snow to drift in these directions. Such drifting might produce glaciers on the east and north sides with few or none on the south and west, especially with the aid of favorable temperature distribution.

Another problem is presented by the restriction of the large glaciers to the northern area, with decidedly smaller ones at the same and higher altitudes in the southern part of the plateau. Here the present distribution of snow shows no appreciable difference. The assemblage of larger glaciers in the southern half of the area shown on Plate 5 must have resulted in the main, however, from pronounced concentration of snowfall. A suggestion of such distribution is given by the fact that there is generally heavier snowfall nowadays in the Gooseberry basin than farther south, but Mr. Humphrey states that this difference is neither constant nor pronounced.

It is worthy of note that there seems to be no evidence of glaciation on the ranges of high mountains east of the main divide. One of these ranges, the line of mountains east of the Joe's Valley graben (Candland Mountain to East Mountain), is partly shown in the southeastern part of Plate 5, where the pattern of outcrop shows plainly the angular outlines of stream-cut canyons and the absence of rounded amphitheatres. These mountains are in considerable part higher than the main crest in the most strongly glaciated district. The absence of glaciation was not due to altitude, nor to exposure; precipitation seems the only possible

factor that could have caused the difference. Here the evidence of modern snowfall is very suggestive. Mr. Humphrey reports that the snowfall is consistently lighter on the eastern ranges than on the main divide. The storms, coming generally from the southwest, largely expend themselves on the western highland, and less moisture is left for precipitation beyond. Doubtless a similar condition obtained in Pleistocene time.

RELATIONS BETWEEN GLACIATION AND FAULTING

Most of the faulting in the Wasatch Plateau is clearly older than the glacial features. As may be seen on Plate 5, the moraines mask completely the faults of the Joe's Valley zone, and no evidence of displacement in the till itself has been found along any of the faults. There is nothing unusual in this, inasmuch as the faults are the easternmost of the great system of Basin Range fractures, on which the major movements probably took place late in Tertiary time. In parts of the plateau, however, there is evidence of post-glacial faulting, some of which is very striking.

In the southern half of the plateau, on both sides of the crest and roughly parallel to it, is a system of normal faults on which the total movement has been very recent. The flat tops of the plateau and the heads of drainage channels on both sides of the divide have been displaced by faults ranging from a few feet to more than 700 feet in throw. There are also open rifts choked with angular rock debris that testify to recent tensional stress. The slightly modified scarps of these faults and the fresh scars of the rifts would alone suggest extreme recency of movement, but the glacial phenomena afford conclusive proof that the faults are post-Wisconsin in age. The heads of several drainage channels that were modified by ice action, some of them true cirques, have been cut off by faults downthrown on the upstream side whose scarps have dammed the streamways, forming lakes in the flat bottoms of the beheaded cirques. An example of this phenomenon is shown on Plate 4, figure 2.

The effects of very recent fault movement have been observed at many places in the Great Basin (Lawson, 1912; Blackwelder, 1928, p. 303, Fig. 6; Eardley, 1934, p. 393-394, Pl. 72, fig. 2); Gilbert (1890, p. 346-347; 1928, Pl. 14, B) has described the prominent fresh scarps cutting glacial moraines at the western base of the Wasatch Mountains. As to recency, these phenomena compare well with those on the crest of the Wasatch Plateau, but as to structural setting they are different. The fresh scarps in the Great Basin occur at the bases of higher scarps and represent renewed movement along older faults, whereas the surface displacements in the Wasatch Plateau embody the total fault movement, and represent new fractures of post-glacial age.

Other evidence, more obscure, suggesting the displacement of a glaciated surface is present at the head of Flat Canyon, whose peculiar form is described in the first part of this paper and whose possible significance as to older glaciation is discussed farther on. Regardless of the manner in which Flat Canyon was shaped, the sharp drop from its head into the basin of Gooseberry Creek suggests a modified scarp produced by displacement of the valley bottom. Other physiographic evidence, however, too detailed to review here, is difficult to reconcile with the hypothesis of fault displacement, and this interpretation is offered merely as a suggestion. If this scarp does represent displacement of the surface, the movement must have been considerably older than the ones described above, because to the north and south of the head of Flat Canyon the scarp is notably modified by erosion.

DATE OF GLACIATION

EVIDENCE IN THE WASATCH PLATEAU

Both depositional and erosional evidence in the Wasatch Plateau bespeak recency of the glaciation, and it is most likely that the major phenomena described on foregoing pages are of Wisconsin age. The freshness of the morainic material and the virtually unaltered surfaces of the moraines seem to leave any other conclusion untenable. Erosional features likewise suggest a relatively short time interval since the glaciation; the cirques and widened valleys have been only slightly affected by postglacial stream work. Comparison with outlying regions tends to confirm the assignment to the Wisconsin, but it raises questions respecting the entire Pleistocene history of the plateau that deserve brief discussion.

COMPARISON WITH OTHER GLACIATED REGIONS

Several of the mountain districts with which the Wasatch Plateau may be profitably compared give evidence of more than one stage of glaciation, and in areas where the glacial phenomena are well developed, such as the Uinta Mountains (Atwood, 1909; Bradley, 1936, p. 195-196), the San Juan Mountains (Atwood and Mather, 1932) and the Wind River Mountains (Blackwelder, 1915), three stages have been recognized. Correlation between these districts is not entirely certain; it is dependent mainly upon comparison of the degrees to which the moraines have been weathered and eroded, and of the amounts of general erosion in the interglacial stages. It is probable, however, that the three stages in the several districts are the same, and it seems almost certain that the last stage is everywhere equivalent to the Wisconsin stage of the upper Mississippi River Basin. The important point in the present comparison is that the moraines of the older stages show the earlier glaciers to have been nearly

all longer than those of the Wisconsin stage. Especially was there great disparity between the glaciers of the last two stages in the Wasatch and Uinta Mountains, not very far from the Wasatch Plateau. With glaciation of apparent Wisconsin age so clearly manifest in the Wasatch Plateau, it seems strange that there is no clear-cut evidence there of older glaciation.

In the consideration of this question it seems safe to begin by ruling out the possible hypothesis that the moraines in the Wasatch Plateau are actually older than Wisconsin; the evidence already presented so far outweighs facts adducible to the contrary that it seems hardly worth while to debate the matter. This question aside, there seem to be three possibilities: (1) For some reason the Wasatch Plateau may not have been glaciated before Wisconsin time; (2) pre-Wisconsin glaciation in the plateau may not have been so extensive as the Wisconsin; and (3) pre-Wisconsin moraines, originally spread beyond the present limits of the Wisconsin, may have been sufficiently altered and destroyed to be inconspicuous or absent.

As to the first hypothesis, if the plateau was not glaciated at the times of earlier glaciation in the neighboring regions there must have been some marked local difference in physical conditions between the earlier Pleistocene and the Wisconsin stage. The only condition of this sort that would seem to meet requirements is difference in altitude; possibly the Wasatch Plateau had not risen to heights adequate for glaciation before latest Pleistocene time. The latest episodes in the diastrophic history of the plateau have not been accurately dated, and it is impossible to deny this hypothesis absolutely, but the physiographic and structural evidence does not seem to support it. The altitudes of catchment basins necessary for the development of glaciers in the Wasatch Mountains were between 8000 and 9000 feet, and excepting unusually favorable conditions, nearer 9000 feet; in the Uinta Mountains, 9000 to 10,000 feet were necessary (Atwood, 1909, p. 67, 91). The limiting altitudes in the Wasatch Plateau were similar (Pl. 5). The highest catchment basins in the Wasatch Plateau are now between 10,000 and 11,000 feet above sea level. The uplift necessary to account for absence or lesser prominence of earlier Pleistocene glaciers would therefore be not less than 1000 feet. If any such uplift did occur, in the relatively short time between the two latest glacial stages of the neighboring mountains, there should be clear evidence of it in physiographic features such as high level terraces and other discordances of surface features. In general such evidence is lacking, although there are some high level pediment-like plains within the plateau that suggest an episode of uplift; the extent and maturity of regional denuda-

tion to be correlated with the erosion that has dissected these pediments, however, seems excessive for so short a time.

Of course, if the entire differential uplift of the plateau had been steady, recognizable physiographic evidence of its latest stages might not have developed. However, steady uplift at the rate involved (1000 feet or more in one-twelfth to one-twentieth of the Pleistocene [1]) would have put the whole movement, ranging from minima of 4000 feet on the northern margin of the plateau to 7000 feet on the west, in the latter part of the Pleistocene. There is some evidence in the regional volcanic succession to suggest that the elevation of the plateau with respect to the valleys on the west was well under way by mid-Tertiary time. In any event if there has been uplift of 1000 feet or more in the latest part of the Pleistocene, the ensuing erosion has attained far greater maturity than the contemporaneous reduction in the mountains nearby, and clear physiographic evidence is lacking.

As to the second hypothesis, if pre-Wisconsin glaciation was less pronounced than the Wisconsin there is little probability that traces of it have been preserved, and no indirect evidence has been found to support this concept; it can only be reiterated that weaker glaciation in the Wasatch Plateau contemporaneous with the stronger glaciation of the Wasatch Mountains seems anomalous.

The third hypothesis, that older moraines have existed in the Wasatch Plateau, but have been largely destroyed, finds some support in evidence which, although far from conclusive, yet must be considered. Most of the larger moraines terminate in wide, flat-bottomed valleys that may well have been shaped by earlier glaciers. Especially striking is the development of such wide valleys in the upper courses of Gooseberry and Huntington creeks, described in the first part of this paper. Gooseberry Valley, in which the Gooseberry moraine is situated, might well be in part an alluviated fault-block valley, but the western part of it has clearly been shaped by erosion and can not be due to faulting alone. Perhaps the outstanding example of this phenomenon is to be found in Flat Canyon and adjacent parts of Huntington Canyon; Flat Canyon is broadly U-shaped and is not the type of broad valley normally shaped by lateral cutting of streams. The upper course of Huntington Creek is similar in cross profile, but it changes downstream to a rocky V-shaped mountain gorge, a condition difficult to explain by stream-cutting alone, but perfectly normal if the upstream part was glaciated.

The valleys occupied by moraines in the Joe's Valley graben likewise are broad and flat below the ends of the moraines. Here the most striking

[1] Correlating the older stage of the Wasatch Mountains with the Iowan (Blackwelder, 1931, p. 918) and accepting Kay's (1931, p. 465-466) estimates for the duration of the Pleistocene.

example is the valley of Lowry Fork, into which the southern of the three piedmont moraines extends; this valley is almost perfectly U-shaped. This profile may have resulted from the coalescence of talus slopes on the opposing scarps of the faults that bound the valley, but it might equally well have resulted from the work of a large glacier. The broad valley north of the central moraine is shown on Plate 2, figure 2.

If these broad valleys were formed by glaciers, however, the moraines have disappeared, and considering the prominence of the older moraines in the Wasatch and Uinta Mountains, this fact considerably weakens the hypothesis of earlier glaciation. Such disparity might be explained by the destructible nature of the bedrock materials in the Wasatch Plateau, as compared with the more resistant rocks of the Wasatch and Uinta Mountains, but even this difference hardly seems adequate to account for the complete destruction of moraines as large as the terminal deposits of the supposed glaciers would have been. There are gravel deposits in both Huntington Canyon and Joe's Valley that might have been glacial in origin, but, as far as they have been studied, might equally well be coarse stream rubble. The deposits in Huntington Canyon are in part terraced and must have been to some extent stream-worked, but those in Joe's Valley, in the lower course of Lowry Fork south of the Olsen Ranch, could easily pass as the much reduced lateral moraine of a glacier. The absence of moraines might suggest even greater age than that of the early stage in the Wasatch Mountains, perhaps comparable to the Little Dry stage of the Uinta Mountains (Bradley, 1936) or the Cerro stage of the San Juan Mountains (Atwood and Mather, 1932) or the Sherwin stage (Blackwelder, 1931) and the El Portal and Glacier Point stages (Matthes, 1930) of the Sierra Nevada, but the amount of stream work later than the broad valleys in the Wasatch Plateau seems too small to stand comparison with the profound erosion that has occurred in the other regions. Summing up the available evidence, it seems fairly probable that older glaciation occurred in the Wasatch Plateau, but none of the features on the basis of which it might be postulated agrees well with comparable features in other glaciated areas.

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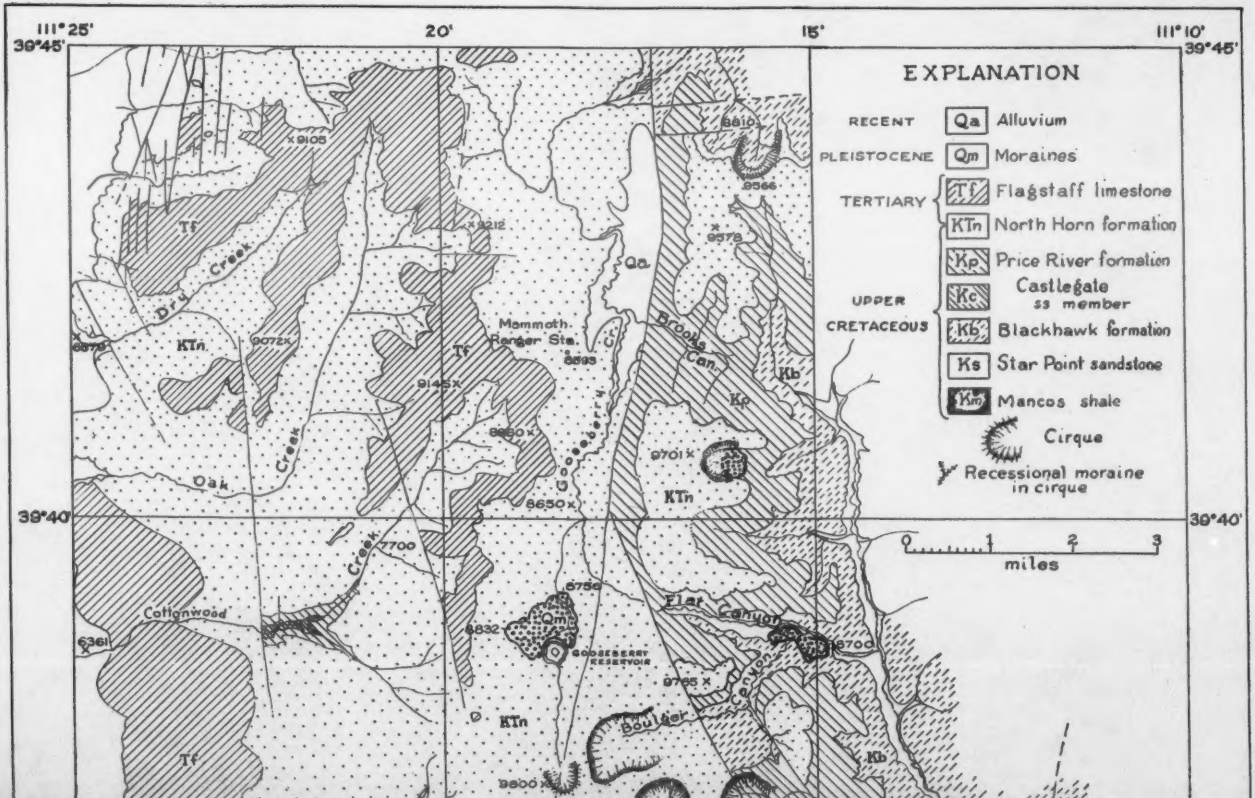
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EXPLANATION
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KTn North Horn formation
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Ks Star Point sandstone
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Lake Canyon
Rolfson Canyon
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Birch Creek
North Creek
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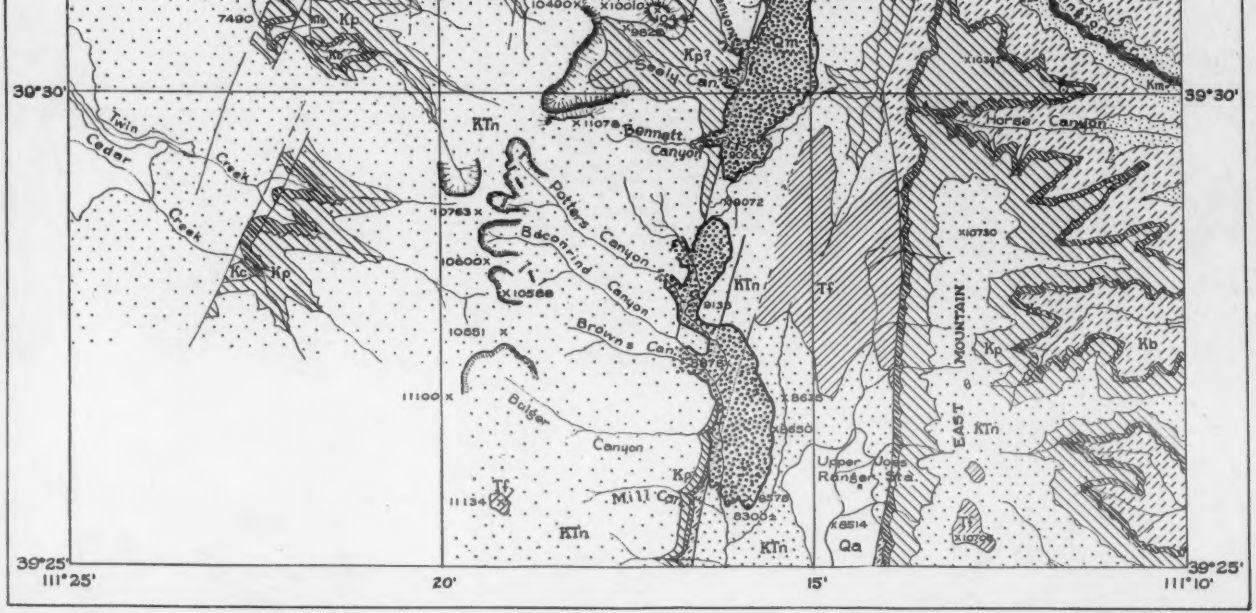

GEOLOGIC MAP OF PART OF THE WASATCH PLATEAU, UTAH, SHOWING GLACIAL FEATURES

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QUATERNARY HISTORY OF LOUISIANA

BY RICHARD JOEL RUSSELL

CONTENTS

	Page
Abstract	1199
Introduction and acknowledgments	1200
Geologic observations	1201
Modern deltas of the Mississippi River	1201
General statement	1201
Teche-Mississippi	1202
Plaquemine-Mississippi	1206
Lafourche-Mississippi	1208
Latest Mississippi courses	1209
Recent stratigraphy	1211
Pleistocene stratigraphy	1213
Topography	1216
Interpretation	1219
Fluctuations in sea-level	1219
Regional tilting	1223
Conclusions	1227
Isostatic implications	1227
Structural implications	1229
Works to which reference is made	1230

ILLUSTRATIONS

Figure	Page
1. Teche-Mississippi course	1203
2. Plaquemine- and Lafourche-Mississippi courses	1207
3. Latest Mississippi courses	1210
4. Centers of Recent, Prairie and Montgomery deposition	1216

ABSTRACT

Five Pleistocene glaciations lowered seal levels sufficiently to rejuvenate the Mississippi River system, enabling it to cut valley systems in Louisiana to depths of 100 to over 300 feet. Rising seas of waning glaciation permitted alluvial drowning of these valley systems and the deposition of five Quaternary formations, the latest being the Recent.

Regional tilting, caused by subsidence of the deltas and a compensating elevation inland, has preserved the surfaces of the four Pleistocene formations as terraces, which extend continuously both coastwise and inland along major streams. In central Louisiana the oldest surface slopes at 10 feet per mile, younger surfaces at rates of about 6, 2, and one feet. The Recent surface slopes less than one-half foot. Slopes of terraces flatten inland and intervals are more constant in northern Louisiana. Slopes steepen coastward, particularly toward later deltas, and each

formation passes below more recent deposits. Sediments grow finer upward in each formation.

Basal gravels of each formation have been identified under the Recent deltaic plain and positions of maximum deposition have been located for four of the five Quaternary formations. Each of these positions is characterized by lenticular thickening downward, so that at least three of the formations attain a maximum thickness of about 3000 feet. Each lens marks the position of the delta of its period. No two center in exactly the same place.

Seven subdeltas of the latest Recent have been identified, heaviest deposition being centered east of a line southward from Baton Rouge. North about 100 miles, in southwestern Mississippi, is the highest land of the lower Mississippi Valley, the result of subsidence of the Recent delta.

INTRODUCTION AND ACKNOWLEDGMENTS

Were the doctrine of uniformitarianism in need of factual support it would be difficult to find any region more admirably suited to the purpose than that of the Gulf Coast Geosyncline. Here we may see today, in active operation, the geologic processes which have been dominant since the late Mesozoic in forming a sedimentary section some 30,000 feet thick, extending coastwise for over a thousand miles, with a width, in its landward limb, at places exceeding 350 miles.

Probably no region of comparable area has been so thoroughly explored beneath the surface. More than 100,000 petroleum wells have penetrated this sedimentary section. In geographical distribution they have sampled its parts thoroughly. No other region has so many deep wells. No other region has such a wealth of drillers' logs, electrical logs, cores, and cuttings available to study.

The story revealed by this unrivaled wealth of geologic information is one of continuous sedimentation, subsidence of coastal territory, and uplift inland. A progressive shifting of shoreline positions southward during the Cenozoic has been punctuated by minor advances and retreats, and modified in accordance with the shifting of delta sites. The division of this stratigraphic section into formational units has been essentially paleontologic in its basis and not altogether satisfactory from the standpoint of either theory or practice. Inland deposits consist of the materials of floodplains and deltaic coastal plains. Sequential gradation leads downdip through brackish water deposits into strictly marine facies. Essentially lacking in unconformities, revealing a diastrophic history duplicating that going on today, and a record of sedimentation having few variants from that of the present, the Gulf Coast Geosyncline region presents an ideal problem in which an understanding of the most recent episodes of geologic history is necessary for a clear interpretation of the past. The following summary of the Quaternary history of Louisiana is presented from this viewpoint.

Louisiana is the theater of most active geologic change along the Gulf Coast Geosyncline today and has been during most of the Quaternary.

For this reason the later episodes of its geologic history are especially clear.

It is evident in the following text that the writer is deeply indebted to his colleagues in the School of Geology of Louisiana State University not only for factual material but for many conclusions as well. Any discussion of the Quaternary of the Gulf Coast must draw upon the deltaic coastal plain concept of Barton (1930) and credit Doering (1935) with a clear discussion of regional tilting.

GEOLOGIC OBSERVATIONS

MODERN DELTAS OF THE MISSISSIPPI RIVER

General statement.—In the Louisiana portion of the alluvial valley of the Mississippi River are the remnants of several main channels which today reveal more or less complete distributary systems closely resembling those of today's delta. The relative ages of these old channels and their deltas are clearly indicated by such criteria as degree of channel abandonment and deterioration, stream patterns, natural levee characteristics, and other techniques of coastal plain geomorphology. Conclusions thus derived are substantiated by means of reference to an Indian chronology which has been established chiefly upon entirely different lines of evidence.

It is inappropriate that these latest deltas of the Mississippi River be referred to the Recent without some comment concerning the meaning of that term and the relative proportion of Recent time involved in their origin.

Following a suggestion made by Reade (1872, p. 118-19), the writer has come to the conclusion that the most satisfactory definition of Recent is that period of time during which the last major rise in general sea level has occurred. Evidence of this rise appears to be world-wide and thoroughly established. The definition, as stated, is divorced from the hypothesis that its underlying cause was a significant retreat of continental ice cover, although little doubt exists that such was the case. The definition implies stratigraphic relationships which conform to accepted usage in other parts of the time scale. The Recent will continue only if the seas continue to rise, or at least to maintain approximately their present levels. If the ice-caps melt completely, Recent strandlines may attain elevations well above the 100-foot contours of today (Daly, 1929, p. 722).

Various estimates, ranging from less than one mm. to as high as 914 m., have been made as to the lowering of general sea levels during times of maximum ice cover (Daly, 1925, p. 285). Antevs (1928, p. 81) summarizes these studies and concludes that lowering amounted to from

290 to 305 feet. It appears that most modern students accept some such value. Louisiana evidence agrees with it.

Recent, as here defined, possibly covers a period of some 20,000 to 30,000 years (Daly, 1929, p. 725; Penck and Brückner, p. 1168; Soergel, 1925, p. 266). During these years the Mississippi River, confronted by a rising base level, has been alluviating a valley cut originally during the lowering base level of the last ice advance. Between the walls of this old valley the alluvium is now not over 300 feet thick. Beyond the walls, fronting the Gulf of Mexico, the surface of the alluvium continues as a deltaic coastal plain between 20 and 100 miles wide, with a depth of alluvium in excess of 3000 feet.

There has been an ever-changing geography during the 20,000 to 30,000 years of Recent alluviation. The flood plain and deltaic coastal plain surfaces initiated at the start of alluviation now lie buried below subsequent Recent deposits. Any delta now evident at the surface, identifiable in terms of stream or natural levee patterns, must have been formed well toward the present day and should be considered modern, rather than simply Recent. Indian chronology suggests an interval of less than 2000 years as the time involved for the building of modern deltas.

Teche-Mississippi.—The oldest of the modern deltas now, or likely to be, identified on the surface was built by a Mississippi River flowing along a channel that had almost nothing in common with that of the existing river across Louisiana. The major relict channel of this abandoned course is now called Bayou Teche in southern Louisiana and the most appropriate name for the stream which created the course is Teche-Mississippi.

The Teche-Mississippi (Fig. 1) south of Red River followed a route which is today continuous in spite of the fact that it bears the names of Bayous Jack, Rouge, Negro Foot, Wauksha, Courtableau, Teche, Boeuf, L'Ourse, and Black (Howe and Moresi, 1933, p. 31-37). North of Red River, in Tensas Basin, the channel continues as Black and Tensas rivers. In the vicinity of the Red the channel is obscured for a short distance by blanketing Red River alluvium.

Numerous Mississippi River channel scars on the floodplain of Tensas Basin indicate a significant period during which the river wandered rather freely. The channels were located mainly near the western valley wall, though some were in the central part of the alluvial valley (Russell, 1933). In sharp contrast, the more southerly Atchafalaya Basin is free from Mississippi channel scars (Russell, 1938a, p. 27-33), suggest-

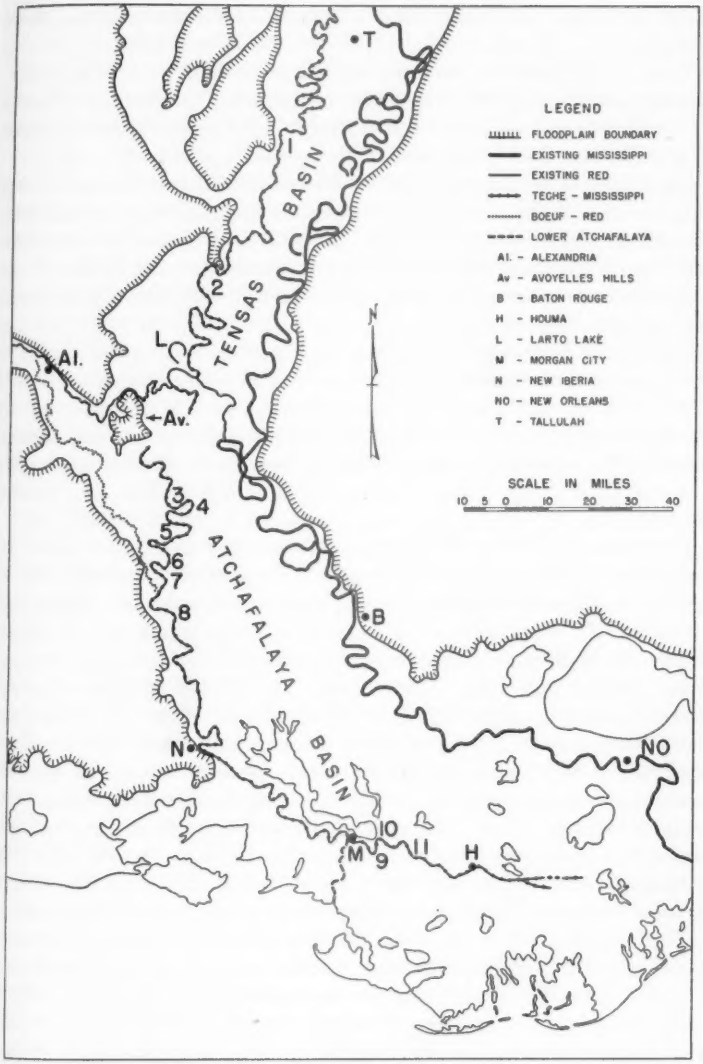

FIGURE 1.—*Teche-Mississippi Course*

Existing channels are called: (1) Tensas, (2) Black, (3) Jack, (4) Rouge, (5) Negro Foot, (6) Wauksha, (7) Courtableau, (8) Teche, (9) Boeuf, (10) L'Ourse, (11) Black.

ing rather close confinement at all times to the Teche channel, along the western valley wall.

Valley wall meander scars of such size and radii of curvature that impingement of only the Mississippi could have left them characterize the bluff lands west of Tensas Basin (Huner, 1939, p. 23-25) and also west of Atchafalaya Basin (Howe and Moresi, 1933, p. 14-15).

During Teche-Mississippi time an arm of the Gulf of Mexico covered the vicinity of New Orleans and extended some distance northward along the eastern side of the present alluvial valley. This area is today characterized by the presence of marine and brackish water shells not far below the surface (Forshey, 1846; Riddell, 1846; Hilgard, 1870; Russell, 1936, p. 148-58).

The Teche-Mississippi resembled the existing river with regard to the orientation of its delta. Just as the river now turns eastward a short distance south of Baton Rouge and, in the main, is characterized by distributaries tending to project eastward, the more ancient channel turned toward the east in the vicinity of New Iberia and formed a delta whose main channels, insofar as they have been identified, also pointed eastward.

In tracing the Teche-Mississippi, or any other old river course, the most useful single criterion is natural levee pattern (Russell, 1939, p. 1212). Channels of abandoned streams often deteriorate so rapidly that a few decades may witness the erasure of every indication of original width. Wide alluvial levees, however, remain for long periods of time. Subsequent alluvial burial or subsidence below levels of adjacent marsh or swamp are the agencies chiefly responsible for levee obscuration and these are relatively slow, particularly some distance back from the coast. Natural levees preserve channel patterns. Streams engaged in aggradational meandering gradually attain a climax pattern complexity, with radii of meander loops developed apparently according to some ratio between volume and stage variation (Russell, 1936, p. 121-37; 1939, p. 1202-10). The natural levees of such streams are as certain evidence of identity as the whorls, loops, and targets of the human fingerprint.

The identification of the Teche-Mississippi channel between northern Louisiana and its delta head, near Houma, is comparatively simple and may be followed readily enough on any reasonably accurate map of the state, using only the criterion of meander pattern. In the field one finds this whole course flanked by natural levees of a width, cross section, and material characteristic of only the Mississippi in Louisiana. West of the Atchafalaya Basin the levees are complicated by the effects of Red River sedimentation within the Teche-Mississippi course.

Red River appears to have followed its own direct course to the Gulf

during much of modern times. Its existing channel across Avoyelles Hills and debouchement into the Mississippi is a very late development (Fisk, 1939b). Older courses lay to the west and are still evident in terms of channel scars. Among these is an extremely interesting channel which developed in post-Teche-Mississippi times, the Boeuf-Red.

The Boeuf-Red (Fig. 1) followed a route southward from Alexandria and into the lower part of the Teche-Mississippi. Its alluvial deposit built a sharply crested pair of red natural levees within the broad gray levees of Mississippi origin. Along Bayou Teche today these double crested levees are very conspicuous, accentuated as they are by shape, color, and soil contrasts. Houses and roads are built both along Mississippi and Red river crests. Ponds and minor drainage channels develop between, forming one of the most characteristic features of landscapes along one of Louisiana's best known and earliest settled bayous. When the Red managed to cross Avoyelles Hills it abandoned the Teche-Mississippi lower course but between the hills and Black River adopted another part of the old channel and reversed its original direction of flow.

The Teche-Mississippi channel is now crossed, at Morgan City, by the main outlet of Atchafalaya Basin. The water of the modern Teche is here diverted and flows to the Gulf through the Lower Atchafalaya River. Double crested levees, however, continue eastward across the break.

A second outlet of Atchafalaya Basin is called Bayou Boeuf and is located several miles east of Morgan City. This outlet has in part adopted a portion of the Teche-Mississippi and its vigorous erosion, chiefly during the last few decades as the result of artificial control, has removed the double-crest levees for a short distance. Farther eastward, along quiet Bayou L'Ourse, the double crested levees are still intact, with red soils confined between gray barriers. For the greater part of its length the main road along this bayou follows the crest of the right bank levee of the Red River. Almost all of the houses here are located on the right bank Mississippi levee and are thus separated from the road by ponds and marshes ordinarily crossed on foot bridges, but the presence of boats at nearly all houses is necessitated by frequent flooding of the low area between levee crests. Beyond Bayou L'Ourse the Teche-Mississippi continues as Bayou Black.

East of Houma the Teche-Mississippi delta has been covered to a very great extent by more recent deposits. Most of these accumulated during Lafourche-Mississippi time, when distributary channels from the north were, in the main, directed southward so that they crossed Teche-Mississippi channels at high angles. Several ridges and channels radi-

ating in easterly directions from Houma are the only surficial clues of the pattern of Teche-Mississippi distributaries. It should be possible to establish the old channels by means of bore holes through the overlying alluvial blanket and thus reconstruct the delta in detail. This is a project the writer expects to undertake in the near future.

The diversion of the Mississippi from its Teche course was one of the most eventful changes undergone by that river in modern times. Future investigation may identify the exact point at which this diversion occurred. At present we know only that it was north of Tallulah, in northern Louisiana. The Tensas channel has been complicated at its northern end as the result of vicissitudes both in Mississippi and Arkansas channel positions. Several crevasse distributaries must be separated from major channels, and history below an alluvial cover must be investigated before the exact point can be fixed. The diversion resulted in abandonment of the Teche-Mississippi channel and the establshment of a route closely parallel to that today along the common border of the states of Mississippi and Louisiana, near the foot of the bluffs along the eastern side of the alluvial valley.

When the Mississippi abandoned its Teche course alluviation had raised the western and central parts of its valley so that they stood relatively high and dry. The eastern part of the valley then consisted chiefly of extensive back-swamp lowlands which offered a very favorable route to the Gulf. Open water and marsh at approximate sea level extended a considerable distance north of New Orleans. The new route offered a steeper gradient, because of lessened distance between point of diversion and sea level and also because at first the channel no longer faced the necessity of following a route lengthened by complex meandering. In time these advantages disappeared. Alluviation resulted in new lines of levees. Meandering resulted in courses having increasing complexity and, ultimately, the eastern course lengthened so that the distance to tidal water exceeded the length of the old Teche-Mississippi itself.

Plaquemine-Mississippi.—The first eastern channel of the modern Mississippi followed the vicinity of the valley wall several miles beyond Baton Rouge. Instead of turning eastward, as does today's channel, about 10 miles south of Baton Rouge, the original route continued southward to the Gulf. The town of Plaquemine marks the point of departure between old and new courses (Fig. 2). Today the rather insignificant bayous, Plaquemine and Jacob, outline the sides of the old channel as far as Crescent (Russell, 1938a, p. 11-16). Below Crescent it may be followed for several miles on the evidence of broad natural levees and for many miles more on the evidence of occasional stream patterns which,

although existing on the surface of more recent swamp and marsh areas, by their shapes and sizes reveal inheritance from the Mississippi itself. The longitudinal gradients of natural levees are steep along the Plaquemine course and 30 miles below Plaquemine, crests pass below the level of younger deposits.

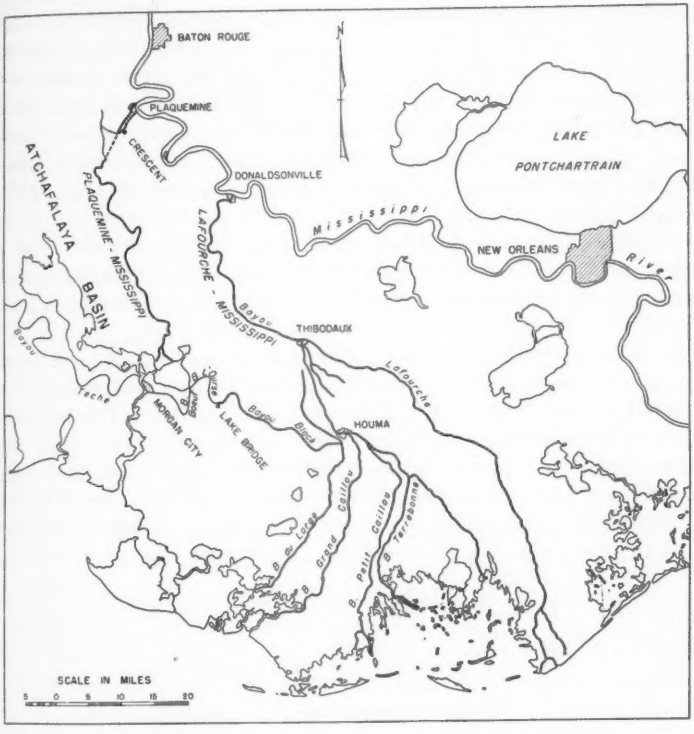

FIGURE 2.—*Plaquemine and Lafourche-Mississippi Courses*

An Indian chronology recently established for the lower Mississippi Valley (Ford, 1936; Kniffen, 1936) appears to be useful in dating the Plaquemine-Mississippi and all subsequent channels.

In applying archaeological conclusions to problems of floodplain geomorphology the basic assumption has to be made that Indians lived most numerously along the active channels of their day. The writer and F. B. Kniffen have subjected this assumption to two rather severe tests. In the first case, each worked out independently the sequence of

channels in the Lower Delta region, south and east of New Orleans. The evidence on one hand was based upon such physical features as stream patterns, natural levee gradients, and stratigraphy, while, on the other, it was based upon potsherd designs, artifacts, and the character of mounds and village sites. The agreement was perfect, although it involved the recognition of an important channel system, the Forts subdelta, which was wholly unsuspected when the work was initiated and which is not particularly evident on the surface today (Kniffen, 1936, p. 417-22; Russell, 1936, p. 111-16). The second test was conducted in Ascension and Iberville Parishes, a surface predominantly swamp today. Field work and writing were conducted independently, conclusions being compared only after the entire work was finished and, again, perfect agreement was found with regard to dating channel sequence (Kniffen, 1938; Russell, 1938a), although, for the second time, the sequence turned out to be one unsuspected when field work was initiated. A similar test is now under way in central Louisiana, where the geomorphological conclusions of H. N. Fisk will be proved by the archaeological results of J. A. Ford.

The Teche-Mississippi, from negative evidence, appears to have flowed in pre-Indian times. No mounds, village sites, pottery, or artifacts have been found which indicate Indian life along the sides of the Teche-Mississippi. The Plaquemine-Mississippi bears mounds and villages of a very old culture, closely related to the Marksville, the oldest yet established in Louisiana and supposedly the equivalent of Hopewell in the Ohio Valley. All subsequent courses of the Mississippi are flanked by sites indicating more recent Indian cultures.

Lafourche-Mississippi.—The third of the modern Mississippi channels departed from the Plaquemine course at the town of Plaquemine and followed the existing Bayou Lafourche (Fig. 2). Natural levees along this bayou are today wider and higher than those along the more modern channel past New Orleans. Whereas the Teche and Plaquemine courses are today distinct enough upstream and somewhat hazy with regard to delta details, the Lafourche-Mississippi exhibits a crystal clear story in both regards.

The original Lafourche-Mississippi delta headed at the town of Thibodaux, not far north of the head of the Teche-Mississippi delta. Here is the primary branching of the main stream into major distributaries, Bayous Black, Terrebonne, and Lafourche.

Bayou Black followed a southerly course as far as Houma, where it encountered the main Teche-Mississippi channel. Unable to cross this barrier, Bayou Black adopted it, reversed its direction of flow, and fol-

lowed it as far as Lake Bridge. The alluviation of Bayou Black buried the old double crested Teche levees and established a steep westward gradient of levee crests. Between Houma and Lake Bridge, in longitudinal profile, levee crests approach the level of water in the existing Bayou Black at a rate of 3 inches per mile. This steep slope is accompanied by narrowing of dry natural levee agricultural lands from a width of more than one mile near Houma to practically nothing at Lake Bridge, 20 miles to the west.

A second "Head of Passes" complicates the pattern of the Lafourche-Mississippi delta. Near Houma, Bayou Terrebonne forks into Bayous Terrebonne, Caillou, and du Large. This is a major subdelta of the Lafourche-Mississippi delta and deserves a name of its own.

The distal portion of the Lafourche-Mississippi delta has been truncated by coastal retreat. Waves have not only removed many miles of delta fingers but have succeeded in driving back a rather continuous beach until it flanks a comparatively smooth coast. The distance from Houma to open water of the Gulf is now slightly more than 32 miles, airline. Originally the distance to the ends of major Lafourche-Mississippi "Passes" may have been over twice this amount.

The natural levees of Lafourche-Mississippi distributaries are now covered for many miles below Houma by oaks and dense vegetation of the "cheniere" type. As the low ridges dip below existing marsh levels the trees disappear abruptly. Between the forested ridges, on widely spaced levees, are marshy fills containing today's small bayous, each with its own little levee system and meander patterns wholly "underfit" with regard to the original levees.

Indian cultures along the original levees of the Lafourche-Mississippi distributaries are younger than those along the Plaquemine-Mississippi, belonging to the Bayou Cutler complex, which was once widespread over coastal Louisiana (Kniffen, 1936).

Latest Mississippi courses.—Diversion of the Mississippi River from its Lafourche to its New Orleans channel at Donaldsonville gave rise to the latest courses of the river. While the effect has been profound with regard to present day population and human activities, this diversion was not as significant from the earth history standpoint as the original abandonment of the Teche-Mississippi channel.

The earliest of the latest courses, the Metairie-Mississippi (Fig. 3), departed from the existing channel at Kenner, about 10 miles above New Orleans, and was directed eastward across the territory now lying between the business district of New Orleans and Lake Pontchartrain. The natural levees are continuous and bear the name Metairie Ridge, upstream.

Farther east they flank Bayou Sauvage, a very decadent remnant channel, finally sinking to marsh level in the vicinity of Chef Menteur. Pearl River alluvial deposits mask the details of the delta itself.

Metairie-Mississippi alluviation terminated the open water conditions

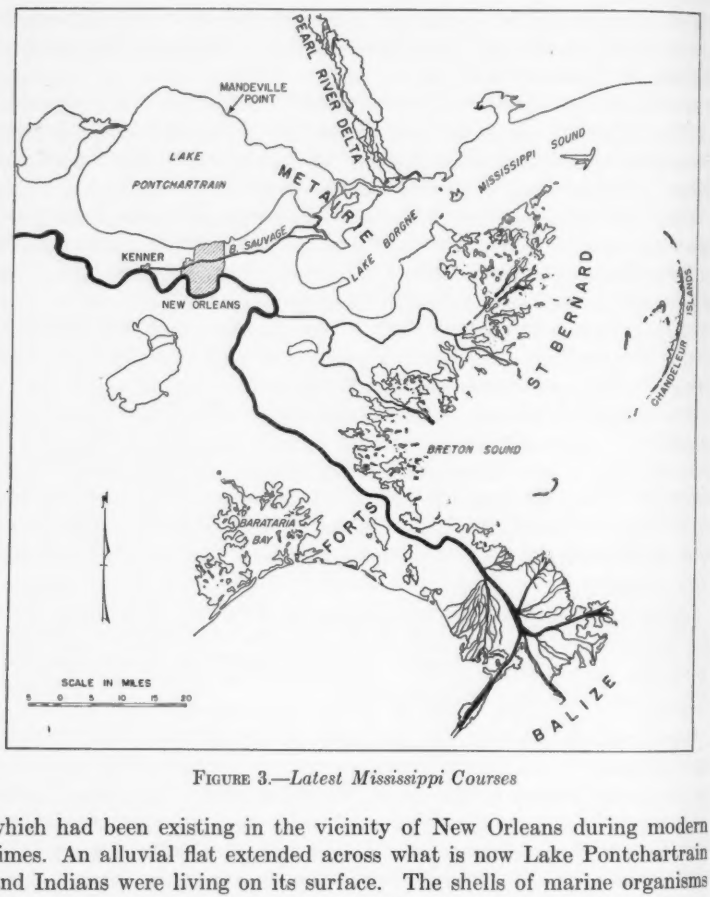

FIGURE 3.—*Latest Mississippi Courses*

which had been existing in the vicinity of New Orleans during modern times. An alluvial flat extended across what is now Lake Pontchartrain and Indians were living on its surface. The shells of marine organisms are now found at depths of about 14 feet in the vicinity of New Orleans. Indian artifacts continue down to at least 10 feet below sea level. In what is now about the deepest part of Lake Pontchartrain, Indian relics are dredged from depths of at least 14 feet. These belong to the relatively late Coles Creek culture.

Latest history takes us to the Lower Delta (Russell, 1936), the region south and southeast of New Orleans. Channel diversion occurred at Kenner. Other diversions farther downstream have resulted in subdeltas, which in order of age have been named Forts, St. Bernard, and Balize. The building of the Lower Delta has been rapid and has been accompanied by subsidence of the region just to the north. The Coles Creek Indian village 5 miles south of Mandeville Point has been lowered to a minimum of 14 feet below sea level so that dredging is necessary to recover its artifacts from the bed of Lake Pontchartrain.

Archaeological evidence is clear in favor of the conclusion that Lake Pontchartrain and the chain of lakes extending west from Mississippi Sound have resulted from downwarping, rather than the alternative explanation that they are residual remnants of the Gulf of Mexico now isolated as the result of Lower Delta alluviation. In the site south of Mandeville Point, recovered materials include Indian bones, animal bones, charcoal, and pottery. It is impossible to believe that materials of such diverse specific gravity were drifted into place and concentrated as a deposit. The assemblage is exactly similar to that of village sites elsewhere and leads one directly to the conclusion that this particular village was inhabited while above sea level. The stratigraphic position of the Coles Creek culture elsewhere is such that identification with Metairie-Mississippi time is very probable. The time suggested for the presence of alluvial flats over the area which is now Lake Pontchartrain is thus the time when a nearby main channel was present to furnish the alluviation.

Only the lower part of the existing Balize subdelta is too recent to bear traces of prehistoric Indian occupance. Other subdeltas exhibit cultures ranging from Bayou Cutler, in the oldest case, to Bayou Petre (Kniffen, 1936), in the St. Bernard. The most modern sites include European trade articles and therefore bring us into historic time.

Archaeologists are as cautious as geologists in supplying time estimates in terms of years. The relative ages of cultures have been established by methods which are in part stratigraphic and which appear perfectly sound (Ford, 1936). Kniffen, with considerable reluctance, ventures the opinion that the entire chronology from Marksville to Historic occupied less than 2000 years. If the Recent is some 20,000 to 30,000 years old, it seems that this archaeological estimate is geologically acceptable.

RECENT STRATIGRAPHY

In its broadest outlines, Recent stratigraphy is concerned with deposits since the last major rise in general sea levels was initiated. Valleys cut during the latest (Wisconsin-Würm?) glacial low-level seas have been

filled to some extent and deltaic plains have been built along the coast. The alluvial deposits are strictly fluviatile upstream, where they lie between flanking valley walls, become transitional downstream, and merge into strictly marine beds in the vicinity of the shore of the Gulf of Mexico.

The history of modern deltas is not only the last chapter in this history but it must be understood in order that the structural relationships of other Recent deposits be clarified.

The Recent formation is apparently the equivalent of the European Flandrienne, including the Ostende, Calais, and Dunkerque (Dubois, 1925, p. 865).

The thickness and shape of the Recent formation is positively determinable from bore-hole evidence. The staff of the Louisiana Geological Survey has engaged on this problem, using hundreds of cores supplied by the Louisiana Highway Commission, the Mississippi River Commission, and various oil companies. They have also taken their own cores at points of particular interest. Hundreds of logs of water wells have also come under inspection. While conclusions are far from final at present, we can advance certain generalizations with utmost confidence.

Between valley walls, some distance back from the coast, the Recent nowhere is over 300 feet thick. Near the junction of the Mississippi and Red rivers it is approximately 250 feet. Along the Louisiana-Arkansas line, Mississippi alluvium is about 175 feet thick. Depths diminish upstream, particularly along smaller streams. The floor of the valley below the alluvium was not particularly flat, so that all values are minima. We have not yet reached such refinements as identification of original channel positions, or deepest thalweg holes. The values above are general and are not isolated observations selected because of any unusual feature.

It is characteristic that the Recent formation consists chiefly of coarser basal and finer upper sediments. Sand and gravel is almost universally found just above the contact with bedrock, or Pleistocene materials, and its amount decreases upward. Any transverse cross section through the Recent shows many lenticular irregularities with regard to grain size but silt and clay sizes predominate only above the basal part of the formation. These characteristics continue coastward, beyond valley walls, but depths to gravel increase enormously.

John W. Frink's study of the Quaternary gravels of coastal Louisiana has yielded wholly unsuspected results. When initiated it was thought that it might be possible to trace the gravels known to occur toward the base of each Quaternary formation from surface outcrops to positions well south in the deltaic coastal plain. This was done but the surprising

conclusion of the study was a demonstration that the actual shape of the Recent and the approximate shapes of later Pleistocene formations could be determined with considerable accuracy. Frink's paper will appear in one of the Bulletins of the Louisiana Geological Survey in 1940 and the writer intends here to draw only upon such parts of it as are absolutely necessary for the present discussion.

It was found that the basal gravels of the Recent dip gently and uniformly along the Mississippi Valley, where the alluvium is confined between walls. On reaching the deltaic plain, the dips increase abruptly and lead into a basin-like structure which immediately underlies the position of modern Mississippi River deltas and which has a vertical thickness of about 3000 feet. The Recent gravels thus have a shape not unlike that of a huge ladle, whose handle extends up the alluvial valley and whose bowl underlies the site of heaviest deposition within modern times.

Frink's work goes far toward settling the long argument between followers of Lyell (1847, p. 118-19; 1872, p. 457), on one hand, and Hilgard (1870; 1912), on the other, concerning the thickness of alluvium in the Lower Delta. Lyell examined materials from a 600-foot well, considered it wholly alluvial, and concluded that the

> "recent deposit if it could be gauged would prove to be far more than 600 feet thick, and might even attain twice or thrice that thickness."

Hilgard believed that marine shells about 30 feet below the surface in the vicinity of New Orleans were Tertiary in age and that even the bed of the Mississippi cut Tertiary deposits across the entire delta region.

Extending Louisiana evidence we should picture the Recent formation of the Gulf Coast as a highly irregular solid. Each major river, such as the Mississippi and Rio Grande, presumably contributes a deposit having the shape of a ladle, with handle up-valley and bowl beneath the deltaic coastal plain. Thin offshoots extend from the handle up tributary valleys. Each minor river and stream entering the Gulf probably forms a lesser ladle. The bowls of all ladles are localized downbulges in a coastwise prism with highly irregular details, lenticular in cross section, thinning under the Gulf to a feather edge whose position is determined not only by the distance to which sediments from the land are carried before final deposition but also by the progressive advance of strandlines over areas formerly land as a consequence of rising sea levels.

PLEISTOCENE STRATIGRAPHY

The identification of four Pleistocene cycles of deposition by Fisk (1938, p. 149-72) has furnished the key for the understanding of pre-Recent stratigraphy. Practically all earlier attempts to interpret the Pleistocene involved the recognition of two major formational units, the older

consisting very largely of sand and gravel, being called Orange Sand, Lafayette, Citronelle, etc., and the younger, consisting of clays, silts, and sands, being called Columbian, Port Hudson, Beaumont, Biloxi, etc. This concept led to all sorts of stratigraphic difficulties but it may be difficult to displace for the reasons that it is well established and superficially appears to be well founded in fact. It is true that the older Pleistocene formations are more gravelly and sandy than the younger but it is also true that each of the four formations is more gravelly and sandy toward the base than upward in its own section. One of the most common stratigraphic mistakes resulting from the older concept was that of correlating basal beds of younger Pleistocene formations with the whole or upper parts, of the older.

An additional complication in Pleistocene stratigraphy arose from the attempt to fix an age and stratigraphic position for the loess of the lower Mississippi Valley, something felt necessary in support of an aeolian theory of its origin. We are now confident, beyond all doubt, that the loess is not aeolian. It has developed out of alluvial silts which now occur near and on bluffs, or in other positions providing an adequate drainage. The great bulk of the loess has been derived from the silts of the two younger Pleistocene formations, but incipient loess occurs in the Recent and some loess is found in the older Pleistocene. Conditions favorable for leaching are necessary to transform silt to loess.

In central Louisiana, Fisk (1938, p. 149-72) found that the Pleistocene consists of four separate stratigraphic units. Each of these is deposited disconformably on some earlier formation, Tertiary or older Pleistocene. Each stratigraphic unit exhibits the gradation from coarser basal to finer upper materials. Basal gravels are characteristic of each formation. Silty surfaces are also the rule. Within each formation are depositional irregularities such as are found in any other fluviatile deposits originating under similar circumstances. The upper surface of each formation is a terrace from the topographic standpoint.

More recently Fisk (1939a) has demonstrated the continuity between the fluviatile Pleistocene terraces of central Louisiana and their equivalent interstream deltaic coastal plains. The mapping of both fluviatile and deltaic elements of these surfaces is now well advanced and thoroughly substantiates Fisk's conclusions not only in Louisiana but eastward along the Gulf Coast as well.

Though in his original publication Fisk did not specifically give each of the four Pleistocene stratigraphic units a formational name, each is in reality a separate formation and should be identified as such. It would be most appropriate to use Fisk's terrace names, Prairie (youngest), Montgomery, Bentley, and Williana (oldest), as formational names.

It is undoubtedly significant that the Pleistocene of Louisiana, on wholly independent evidence, has been found to consist of exactly the same number of formations as that of Europe (Depéret, 1918, p. 480-86).

Paleontologic identification of the four Louisiana Pleistocene formations is not possible as yet, but a fair start has been made on the problem (Richards, 1938a, 1938b, 1939; Brown, 1938; Steere, 1938). Thus far initial studies have involved both marine and fresh water mollusca, plants, and mosses. Older studies, such as the careful work of Shimek (1902) on loess faunas, the works of Hay (1915) and others on vertebrates, and numerous studies on Pleistocene invertebrates will take on new meaning when placed in the perspective of a stratigraphy now understood. There is probably enough printed information today to permit, when assembled and digested, as clear differentiation of the paleontology of the Gulf Coast Pleistocene into its four formational units as has been made for the Mediterranean shores (Gignoux, 1913; Depéret, 1918) or Black Sea hinterland (Lungershausen, 1938).

The four Louisiana Pleistocene formations are distinguishable downdip below the deltaic coastal plain. The technique of this study involves starting at surface outcrops of gravel and tracing the basal parts of each formation from well to well along numerous sections. Most of the sections thus studied by Frink were almost normal to the direction of the coast. Sections parallel to the coast were used in tying conclusions together in a three-dimensional result.

Youngest Pleistocene (Prairie) gravels indicate a formational shape very similar to that of the Recent (Fig. 4). The bowl of the ladle has approximately the same depth and area but its location is some distance to the west of the Recent bowl. Deduction based upon Recent conditions would lead us to the conclusion that this bowl underlies the deltas formed during Prairie deposition. Confirmation of this conclusion is at hand, for the reason that stream patterns are still evident on the surface of the Prairie terrace. A very detailed study of these channels has been made by Fisk in central Louisiana (1939b). They have also been noted farther toward the coast (Howe and Moresi, 1933, p. 18-23, with aerial photograph, Pl. II; Howe, Russell, and McGuirt, 1935, p. 4-12). The channels are distinctly those of the Mississippi and they lead directly to the region immediately above the bowl found by Frink.

Evidence as to the shape of the next older Pleistocene formation, Montgomery, is not as conclusive. There is a thickening to at least 2400 feet some distance northwest of the Prairie bowl but outlines are not as yet complete. Still more obscure are the shapes of the two earlier formations, Bentley and Williana, though there are rather definite indications that the Bentley exhibits two centers of maximum deposition.

one on either side of Louisiana, and that the Williana has its most pronounced center well to the east of the state.

TOPOGRAPHY

Recent floodplains and deltaic coastal plain account for slightly less than one-half of the area of Louisiana. This region, in its natural con-

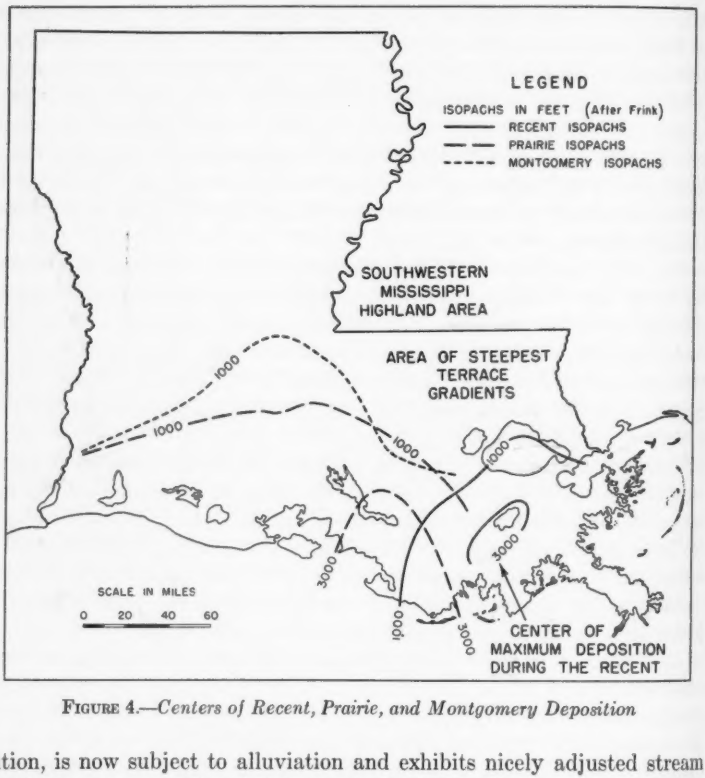

FIGURE 4.—*Centers of Recent, Prairie, and Montgomery Deposition*

dition, is now subject to alluviation and exhibits nicely adjusted stream gradients. In a meridional distance of about 235 miles, from the Arkansas boundary to the Gulf of Mexico, the Mississippi drops about 85 feet, or a little less than 0.32 feet per mile, airline. This is a sufficient approximation for the slope of the major flood plain. The gradient of the river itself, with its complex meandering and several departures from meridional course, is naturally much less but does not concern us here. The flood plains of smaller streams are steeper and in the case of minor

tributaries reach several tens of feet per mile. The coastal marsh is very nearly flat, its chief slopes being those of natural levees, which, in general, are transverse to regional gradients.

In the upland region of Louisiana, Tertiary rocks are exposed at the surface. The Tertiary belt covers most of northern Louisiana. It is crossed by Recent flood plains and by Pleistocene terraces. The terraces are extensive, covering an area about equivalent to that of the exposed Tertiary. Recent valleys tend to follow routes of Pleistocene valleys, so that one or more Pleistocene terraces commonly flank active flood plains. Some of the older Pleistocene terraces reflect flood plain patterns quite unlike those of today. We must await further field work before it is possible to present a detailed picture of earlier drainage patterns.

Some distance south of the Tertiary belt, Pleistocene terraces merge fluviatile characteristics into those of deltaic coastal plains. In this regard they duplicate topographic and depositional patterns of the Recent. The four deltaic plain surfaces of the Pleistocene are separable on the basis of differing gradients, the oldest sloping most steeply. The combined width of all Pleistocene deltaic plains is somewhat greater than the width of the Recent deltaic coastal plain. Many have tried to interpret this region as one of marine terraces but no evidence in support of this thesis is to be found in the field. There is a complete absence of sea cliffs, offshore bars, or other shoreline forms. Only in limited areas of the Prairie formation are marine Pleistocene deposits exposed at the surface today.

The highest points in Louisiana attain elevations of about 500 feet and occur along its northern border. Resistant Tertiary formations form several cuestas in the central and northern parts of the state and these are separated by lowlands along less resistant belts of Tertiary. The dip of the Tertiary is Gulfward in its broader aspects, deltaward in detail. Deltaward dips alter strikes so as to produce the effect of a "Mississippi Embayment," a feature long recognized, but only in a special sense a reality. Valleys cut in Tertiary rocks contain the Pleistocene and Recent deposits of northern Louisiana. Farther south the Tertiary is completely blanketed by more recent deposits.

The four Pleistocene terraces can be differentiated most easily upon the basis of individual gradients. In central Louisiana the Prairie (youngest) terrace is only slightly steeper than modern flood plains. Older terraces, in order of increasing antiquity, exhibit gradients of 1.5, 5.0, and 8.0 feet per mile (Fisk, 1939a, p. 189). Slopes flatten upstream and investigations now suggest that rather constant vertical intervals separate terraces farther inland. When these are thoroughly established it will be interesting to compare the indicated epeirogenic history with

that of western and southern Europe as originally suggested by Boubée (1834, 1847) and Tardy (1877, 1884) and established by de Lamothe (1901, 1915, 1918, 1921), Kilian (1900), Gignoux (1913), Chaput (1917, 1910), Depéret (1894, 1918, 1922), Russo (1920), Roman (1922), Dubois (1925), Brooks (1917), De Martonne (1929), Gallois (1928), Antevs (1929), Anderson (1936), and others.

The topography and areal geology of the transitional belt between the Tertiary area and the belt of Pleistocene deltaic plains is determined chiefly by differing terrace gradients and formational thicknesses. The terrace formations individually vary from about 90 to 300 feet in thickness along master streams and are much thinner along tributaries. Subsequent erosion has removed large quantities of terrace deposit, either paving the way for later terrace deposits or extending the area of Tertiary exposures. All land rising above the level of the highest terrace is composed of Tertiary rocks. The Tertiary is exposed also between the basal gravels of any higher terrace and the top surface of any lower terrace. These Tertiary exposures are wedges which taper and end toward the coast as the steeper gradients of older, upper terraces converge with the gentler gradients of younger, lower terraces. Toward the south, where the terraces expand into their coastal plain elements, Tertiary outcrops are more and more confined to narrow valley wall exposures along existing streams. In southern Rapides Parish, in south-central Louisiana, the southernmost fingers of the Tertiary wedge out completely.

Terrace gradients steepen southward along well marked zones of flexure which traverse the coastal plain element of each Pleistocene surface. Little may be said in this connection about the older Pleistocene surfaces, but the case is clear with regard to the younger. The Prairie (youngest) terrace slopes at a rate of 1.4 feet per mile for the first 10 miles south of Baton Rouge and at 2 feet for the next 10 miles (Russell, 1938a, p. 78). A similar condition obtains west of the river for both Prairie and Montgomery terraces (Fisk, 1939a, p. 191-93, Fig. 3).

Increased terrace gradients observable on the surface are minor in comparison with the changes in slopes of basal gravels. A short distance south and southeast of Baton Rouge basal Prairie gravels dip at a rate of from 100 to 130 feet per mile. A careful examination of the gradients of both terrace surfaces and basal gravels establishes the important generalization that basal gravels dip essentially toward centers of most active deposition at the time when they were formed, whereas terrace gradients slope toward the centers of most active deposition during subsequent stages (Fisk, 1939a, p. 192-99).

Immediately north of the region of modern deltas of the Mississippi River, in the southwestern part of the state of Mississippi, is the highest

land of the lower Mississippi Valley (Fig. 4). Summit areas are here more than 500 feet above sea level, though tide water, in Lake Pontchartrain, lies but 60 miles away. This average gradient, in excess of 8 feet per mile, has no counterpart elsewhere in Louisiana. Terrace gradients are all much steeper here than at other places. The Prairie slopes at a rate of 5 feet per mile, the Montgomery at 8 and more, and older terrace gradients amount to as much as 25 feet per mile. These gradients flatten abruptly in the highest part of this highland and, according to reconnaissance fieldwork by the writer and Fisk during the summer of 1939, slope gently downward toward the north. The highest land in the region appears to be Pleistocene terrace, above which no hills of Tertiary rocks rise.

INTERPRETATION

FLUCTUATIONS IN SEA-LEVEL

As each of the five Quaternary surfaces of Louisiana is underlain by its own geologic formation and its fluviatile element confined to its own system of valleys, there is no escape from the conclusion that five cycles of alternate cutting and filling characterize Quaternary history. It seems most likely that these cycles were caused by major fluctuations in sea level.

As the outstanding difference between Quaternary and time immediately preceding was widespread, cyclic glaciation of continental areas, it is reasonable to suppose that fluctuations in sea level were caused by the extraction of water from the oceans during glacial advance and its return as glaciers waned.

The Quaternary valleys of Louisiana were evidently cut as seas lowered and filled as seas rose.

The quantitative estimates of Daly, Dubois, Antevs, and others concerning the extent to which extraction of water lowered glacial seas agree so closely with depths of fill measured in Louisiana that the author feels justified in believing that these lines of evidence, wholly unrelated in factual background, confirm each other. One is based upon estimated area and thickness of ice cover, the other upon depths to basal gravels in fluviatile terraces. Each suggests a lowering of levels of about 300 feet for maximum glaciation.

If this conclusion be correct, it must be capable of substantiation from other regions because the lowering of sea level was a world-wide phenomenon. This appears to be true but much of the evidence is qualitative at present for the reason that depths of alluvial fill and thicknesses of Pleistocene formations are not as well established for other regions as for the lower Mississippi Valley.

It also appears that continental shelf depths are related to the low level seas (Shepard, 1939). The discovery of "the distinct remains of a delta formation now at about 43 fathoms depth" (Veatch and Smith, 1939, p. 14, 40, 42) at the end of a distinct Hudson River channel, on the edge of the continental shelf, appears to confirm estimates as to the lowering of the seas during the latest glacial advance.

Louisiana evidence has not furnished any clues as to the levels of interglacial seas nor is it likely ever to do so. Coastal subsidence has lowered all older strandlines to positions far below the existing sea level. Their positions are now so far removed from the levels at which they originated that evidence of interglacial sea levels, and for that matter older glacial sea levels as well, no longer exists in the state. It may be possible to find some region far removed from any delta which will yield clues as to interglacial sea levels and thus furnish information relative to the extent of deglaciation experienced at various times during the Quaternary.

The number of high stands of interglacial seas appears to be five, including the Recent, in widely removed regions and thus universal in aspect. Before discussing this matter in detail it is necessary to clarify our terminology.

Geologists are fully cognizant of the fact that both in Louisiana and elsewhere local terrace sequences are complicated by the existence of "fausses terrasses, terrasses irrégulières" and "terrasses secondaires." Many regions exhibit terraces which have resulted from the swinging of a stream from one side of its valley to the other during entrenchment. Terraces also result from changes in gradient caused by shifts in distance to the point of debouchement. Westward flowing streams in the state of Mississippi typically had flood plains adjusted to the Teche-Mississippi course. Such streams were shortened when the Mississippi adopted its present channel and increased gradients have permitted entrenchment, resulting in an excellent system of modern terraces. Benches are also produced along valley sides as the result of lateral corrasion and these become terraces after rejuvenation has become effective. It is not with such features that we are now concerned. It is with the "nappes principales," terraces developed regionally, existing at many places at similar levels on both sides of valleys, and traceable not only upstream along the complex ramifications of tributary systems but also downstream into their broad deltaic coastal plains.

These major surfaces in Europe have been called Flandrienne (Recent), Monastrienne, Tyrrhénienne, Milazzienne, and Silicienne (oldest). The Louisiana equivalents are the Recent, Prairie, Montgomery, Bentley, and Williana. Dubois (1925) correlated the European surfaces with

the McHenry, Talbot, Wicomico, Sunderland, and Lafayette. Scores of other names have been introduced for them and their relative merits may be argued indefinitely, but systems of nomenclature attempting to increase or reduce the number of principal surfaces may well be regarded with suspicion. The writer bases this conclusion not only upon Louisiana studies but also upon field observations along the Rhine, Meuse, and other European rivers, especially the lower Rhone, where he has always been able to identify the same number of principal surfaces.

Particularly confusing in terrace literature are attempts to designate principal surfaces in terms of glacial records. Once the major outlines of the Quaternary had been established (Penck and Brückner), terrace students began using glacial terminology and terraces were usually assigned to times of glacial advance. The underlying reason for this seems to have been an over emphasis of fluvio-glacial deposition. Many terraces were regarded as the result of torrential deposition, others as the result of corrasion, by streams either issuing normally from ice-caps or swollen in volume as the result of ice melting. Terrace surfaces have commonly been regarded as the same age as morainal deposits either upon them or projecting through them.

While it may be true that certain terraces are formed at times of ice advance, the case for the principal surfaces is decidedly opposed to this idea. As Fisk has shown (1938, p. 150-72; 1939a, p. 187), the major Quaternary terraces are depositional and occur in valleys which were cut during periods of lowering seas and which were filled as seas rose again. To the terrace student the concepts "glacial" and "interglacial" are less significant than "waxing" and "waning" of the great ice-caps. Times of maximum ice advance, if prolonged, were times when valleys were being widened at lowered base levels. Times of glacial wane witnessed higher and higher floodplain levels, the culminating stage of which was the time of maximum deglaciation. Alluvial terraces should thus be identified with interglacial rather than with glacial stages. G. E. P. Smith (1938) has advanced similar ideas with regard to the four principal surfaces in Arizona.

Whether Fisk's correlation of Prairie with Peorian, Montgomery with Sangamon, Bentley with Yarmouth, Williana with Aftonian, or the implied correlation in Europe, starting with Sicilienne and Günz-Mindel, will prove correct is relatively immaterial. However it is important that we recognize valley cutting during Günzian ice advance, filling during Günzian wane, as having formed a surface which became a terrace during Mindelian ice advance, and that repetitions of these events are responsible for the major Pleistocene terraces.

Glacial students may oppose the outline here presented on the grounds

of its extreme simplicity. Detailed mapping of glacial deposits certainly suggests a more complex history than is indicated by five cycles of cutting and filling. Daly's Antarctic penguin (1929, p. 721) might well oppose the idea on the basis of exactly contrary evidence. Everyone recognizes that many oscillations between advance and retreat occurred during the major swings of the ice front. The most complex glacial history is bound to occur neither toward the most expanded positions of ice fronts nor toward their retracted positions of today but somewhere in the middle ground. Rejuvenation of streams resulting from lowering seas or filling resulting from rising seas must have taken place whenever minor oscillations were well enough marked to affect general sea levels. Periods of cutting might have been interrupted many times by minor amounts of aggradation. The five cycles represent only the very broadest outlines of glacial history, the major swings from glacial to interglacial conditions.

The main outlines of Quaternary history seem to be more plainly recorded beyond the regions of glaciation than within them. Interpretations of ice advance and retreat from terraces is just one example. The classic studies of Gilbert and I. C. Russell in the lake basins of the arid West is another. Darwin, Dana, Davis, Daly, and others have based important conclusions on evidence from tropical coral reefs, from which have arisen our modern concepts of swinging sea levels related to extent of glaciation. All of these lines of evidence appear to harmonize as to conclusion and furnish a reference frame into which may be fitted the detailed information supplied by students working on deglaciated regions.

The five major low sea-level stages were responsible for an impressive record of erosion. Great valley systems were cut during each rejuvenation. To reconstruct the prisms of material removed from continental areas we must consider each cycle of valley cutting individually. Bore holes must furnish details not of one "deep stage," but of five. We are far enough along on this problem in Louisiana to feel both that this question is soluble and that the volume of transported material is several times greater than that which would be necessary to fill all existing valleys of the Mississippi drainage basin.

From magnitude of Quaternary erosion it is but a step to the complementary problem of locating the deposits formed when seas were low. That they have accumulated mainly beyond emerged continental limits is evident. The bottoms of Pleistocene valleys reach today's shores well below existing sea levels. No suitable mechanism has been suggested for widely dispersing the products of continental erosion over the major ocean basins. Substantial evidence (Schott, 1939) exists that the ocean

floors are free from such material. Thus we must turn to continental shelves as the sites for the bulk of the deposits.

Large areas of continental shelf approximate levels of glacial seas. Departures may be explained on the basis of regional warping. A wide shelf occurs typically at places where Pleistocene drainage patterns indicate probability of wide coastal plain. The distribution of sediments on the shelf is in accord with the idea that deposition took place under low sea level conditions (Shepard, 1939). The amount of bedrock on shelves is probably exaggerated on charts for various reasons suggested by Shepard and also because compact algal growths are likely to be reported as such. Even so, it is not excessive in comparison with bedrock on existing coastal plains and can hardly support a theory of marine abrasion as opposed to a depositional origin for the shelf.

The various suggestions of old shores described by Veatch and Smith (1939), or those which are evident on their charts, probably represent halts during the Recent rise. From evidence on the continental shelf it may eventually be possible to unravel the complicated story of minor fluctuations associated with this rise and thus establish important conclusions with regard to the details of latest ice retreat. Cores and geophysical information will be necessary in supplementing conclusions made on the basis of topography.

In summary, it seems clearly established that Quarternary seas fluctuated in level and many lines of evidence substantiate the conclusion that Louisiana terraces have been formed as a result. The fluctuations in sea level appear to be an excellent guide to the major outlines of glacial history.

The characteristics of Louisiana Pleistocene valleys shed additional light upon the glacial record. Pre-Williana and pre-Bentley valleys (Fisk, 1938; 1939a) were typically wide and had gently sloping walls. This suggests a long, low stand of the sea, or a long stage for each of the first two glaciations. Later valleys have steeper walls, especially the pre-Recent examples, indicating shorter glacial advances. The deepest fill, about 300 feet in valleys, characterizes the Prairie formation, suggesting a greater intensity of glaciation for the pre-Prairie stage than for any other.

REGIONAL TILTING

Terraces resulting from fluctuations in sea level would hardly be identifiable today if all high strandlines approximated the same level. Various amounts of deglaciation might be responsible for differing positions of strands but we should hardly expect an orderly arrangement in which a progressive lowering occurred for each interglacial stage. The French

school of terrace students, who have leaned strongly toward eustatic explanations for terrace sequences, apparently have found this question very much of a puzzle.

It was observed prior to 1834 by Boubée that higher terraces are older than lower, the main evidence being the greater leaching and decomposition observable on higher terraces. These ideas were amplified by him, on the basis of increasingly wide observation, many times in the decades to follow.

Boubée's thesis appears to be universally true. Students everywhere are in rather close agreement that the "nappes principales" occur in an age sequence from youngest at the base to oldest on top. Boubée's evidence is duplicated in Louisiana and has been expressed many times in geologic literature. "Orange Sand" and reddish colors have been associated with older deposits, "buff" and yellowish colors with younger. More scientific conclusions based upon evidence of base exchange and silica-alumina ratio serve better to demonstrate the thesis. Admonitions of Boubée issued a century ago with regard to the greater success of crops on lower terraces have been verified in agricultural experience in Louisiana.

It must be admitted that the evidence of differential leaching alone would not constitute very acceptable proof of relative age in the terrace sequence. Horizontal contrasts are sharp between various parts of a single surface and grain size has much to do with eluvial history. Other lines of evidence clinch the argument.

Stratigraphic proof of terrace ages is ironclad. Each principal surface, when followed Gulfward, converges in slope with the surface next below, passes beneath, and may be followed down-dip as the top of the formation stratigraphically next below the terrace formation crossed. This relationship is naturally freshest and most striking as exhibited between the Prairie terrace and the Recent alluvium. Oxidized subsoils of the Prairie terrace have been traced for considerable distances below the reduced marsh deposits of the Recent deltaic coastal plain (Howe, Russell, and McGuirt, 1935, p. 61-63). Frink has carried every basal Quaternary gravel down-dip from its point of outcrop and has shown that in subsurface stratigraphy the oldest formation underlies the next younger, and so on up to the surface.

Geomorphological proof of greater age for higher terraces is furnished by stream patterns (Fisk, 1938, p. 51-64; Russell, 1939, p. 1222-24). Each terrace has its own type of stream branchwork, dissection being progressively greater in upper terraces. Few streams cross from one terrace to the next in comparison with the density of drainage net on each surface. At many places terraces are separated by "rim-swamp

streams," which hug escarpment bases in exactly the same way that streams today, along the margins of existing floodplains, tend to follow the low ground at the base of the walls of alluvially drowned valleys.

Another excellent measure of age relationships is the increase in terrace gradients from the base to the top of the series. This is most evident in the flexed region immediately inland from the zones where upper surfaces pass below the deposits of next younger formations.

Paleontologic age sequence is being investigated for Louisiana terraces at present. In Europe both paleontologic and achaeologic evidence assists in determining the greater antiquity of upper surfaces.

Various hypotheses have been advanced as to why older terraces occur progressively higher in the Quaternary topographic sequence. Many geologists, especially in France, have looked toward eustatic shifting of sea levels, though few have attempted to present a rational mechanism by which the result has been obtained. Continental depression resulting from ice load has been suggested, as has been the gravitational attraction of ice in raising oceanic waters to unusual heights around the borders of glaciated continents. Modern authorities appear in close agreement that the extraction of water necessary to form icecaps more than compensated either of these tendencies toward higher strandlines (Daly, 1925, 1929). None of these hypotheses is very helpful or convincing in regard to age sequence. The improbability seems patent that eustatic shifts in level occurred in a regular sequence from highest to lowest in terms of age.

In Louisiana it appears clear that the origin of each Quaternary surface is related to a cause which is divorced from the reason it has been preserved in a position where age identification is related to topographic position. Each surface originated as the result of a cycle of eustatic shifting from high through low and back to high sea level. Each surface thus formed has been preserved and differentiated from other surfaces as the result of regional tilting (Doering, 1935; Fisk, 1938, p. 69, 1939a; Russell, 1938b). The recognition of this distinction between cause of surface and reason for existing position at once accounts for the greater antiquity of higher terraces and frees us from such speculative difficulties as the necessity for progressive lowering of each succeeding interglacial strand line.

Regional tilting in a direction which has given greater slopes to terraces in proportion to age is in reality the continuation of a pattern which characterizes the whole Cenozoic history of the Gulf Coastal Plain. Regional dips are steeper in the older Tertiary and more nearly approximate those of the Quaternary in the younger.

In both Tertiary and Quaternary structures it is characteristic that

regional dips steepen along definite zones. In the Tertiary this relationship between up-dip and down-dip portions of the same formation is exhibited along such zones as the Angelina-Caldwell flexure (Veatch, 1906, p. 68; Teas, 1927, p. 681-86; Howe, 1936, p. 19-20; Fisk, 1938, p. 175-77) and Tunica-Mobile flexure (Howe, 1936).

Fisk (1939a, p. 142) has recognized two major types of regions with regard to Quaternary flexing, an epeirogenic region inland, which exhibits only epeirogenic uplift, and a geosynclinal region toward the coast, in which downflexing has been extreme. The zone between these regions consists of epeirogenic and geosynclinal margins. It is here that conditions are transitional and where the sharpest contrasts may be noted between terrace gradients. Central Louisiana measurements such as 8 feet per mile for the slope of the Williana and 5 feet for the Bentley refer to gradients in the geosynclinal margin. Farther inland, in the epeirogenic region, slopes flatten and intervals between terraces remain rather constant. This observation has been checked by the writer and by Fisk along both sides of the Mississippi Valley below Cape Girardeau and agrees with the findings of many European students, especially de Lamothe and his followers.

Interesting hypotheses may be advanced if a rather uniform rate of tilting be assumed. The departures in maximum slopes between the two older terraces are wider than between any of the younger ones; this suggests longer intervals between the earlier than between more recent glacial advances. The very slight departure between Prairie and Recent slopes suggests a very short duration for the pre-Recent ice advance.

The extent to which Louisiana observations with regard to tilting may be generalized in application to other regions is worthy of investigation. Granting that fluctuations in sea level were universal and that all areas affected by shifting base levels underwent alternations between cut and fill, it does not follow by any means that the mechanism for differentiating and preserving surfaces existed everywhere. It is also possible that if such mechanism existed its underlying causes differed from those of the Gulf Coast.

The fact that four "nappes principales" above Recent flood plains seem to exist in many regions far removed from the Gulf Coast is suggestive that regional tilting, or possibly only epeirogenic uplift, has been a common continental experience during the Quaternary. In the delta of the Rhone conditions are very similar to those of Louisiana. The terrace gravels of the Crau pass below the Recent alluvium (Baulig, 1927, p. 501-02) and the same flexing of surfaces occurs along definite zones where older terraces pass below the deposits associated with younger. The writer has also noted this condition along the lower Rhine and feels

that it renders perfectly understandable the stratigraphy implied in the paleontologic correlations of Depéret (1922, pp. 1502-05). The type of area most in need of investigation is one in which major terraces exist along some continental margin far removed from any delta, if one is to be found.

CONCLUSIONS

ISOSTATIC IMPLICATIONS

It seems difficult, if not impossible, to accept Louisiana evidence without coming to the final conclusion that the Gulf Coast Geosyncline is the result of sedimentary loading.

Before attempting proof of this assertion it is necessary to explain the meaning of the term "geosyncline" as used in this connection. The writer will not defend the thesis that it is an exclusive, or even proper, use of the term. He adopts it because it has found its way into literature and common usage.

The writer has elsewhere reviewed in considerable detail the original definition of "geosynclinal" by Dana, the ideas of Hall and others which preceded the definition, and the various hypotheses which have been associated with the term (Russell, 1936, p. 180-93). It is only necessary to repeat here that the definition as originally used referred to large troughs containing sedimentary rock later uplifted into mountainous structures. The mountains themselves were called synclinoria by Dana, a definition which suffered a curious change at the hands of Van Hise (Jones, 1938, p. lx-lxi; Van Hise, 1896). Though an appropriate terminology has not arisen it seems desirable that a distinction should be made between inactive geosynclines of synclinoria and active synclines, such as the Gulf Coast, Po Valley, and Ganges Valley, which are all still in the stage of being alluviated and filled.

The original definition of a geosynclinal (Dana, 1873) appeared in a paper "On some results of the earth's contraction from cooling . . . ," and carried with it a hypothesis concerning origin. The motive force for producing troughs was tangential compression; whether sedimentary filling occurred or not was of secondary importance. The search for unfilled, or partially filled, geosynclines has led some to believe that none exist today and others to look upon oceanic deeps and troughs as examples of unfilled geosynclinal troughs. In common practice the hypothesis of origin has to a great degree been divorced from the concept of a geosyncline, but there is a growing need for a terminology which will distinguish between the troughs of synclinoria, troughs apparently resulting from sedimentary load, downfolds in the floor of the sea, and other similar structures (Hess, 1938, p. 79).

In speaking of the Gulf Coast Geosyncline the writer means a body of sedimentary rocks having great thickness and areal extent, whose basement may be considered trough-like and whose upper surface is essentially horizontal. Its stratigraphy and petrology appear to be essentially those of the Dana geosynclines now exposed in Dana synclinoria. There is little or no evidence in favor of Dana's suggested origin and a body of substantial facts against it.

The sedimentary record of the Gulf Coast Geosyncline indicates no great departure at any point from processes taking place today. Continental, near shore, and shallow-water sediments are typical of its whole Cenozoic column, just as they appear to be in the great sections revealed in older synclinoria. Depression of the floor of the "trough" appears to have been so perfectly adjusted to rate of "filling" that attempts to interpret the growth of the structure as due to anything but sedimentary loading appear far-fetched.

The competence of sedimentary loading to depress material beneath and thus permit additional load at the original site seems clearly established by the lithologic and structural record. The ladle-like shapes of Recent and latest Pleistocene formations of the Gulf Coast should be evidence enough. The location of the "bowls" of the ladles under the places where stream patterns show that deltas existed is critical evidence, most certainly demonstrable in the Recent and quite certainly in the Prairie. The ease with which vertical yielding of the crust takes place appears to be established with reference to ice load during the Pleistocene (Jamieson, 1865; Shaler, 1895; Nansen, 1922). Even the loading of water in the Bonneville basin may have depressed the crust locally (Jamieson, 1887).

While loading is taking place a delta appears not only competent to depress the rocks below but it also drags down a considerable amount of the surrounding territory. These lowered regions, marginal to deltas, have been called "delta-flank depressions" and they seem to characterize the major deltas of the world (Russell, 1936, p. 7-9). Open water occurs in many delta-flank depressions where subsidence exceeds sedimentation. The western end of Mississippi Sound and the chain of lakes extending through Pontchartrain to Maurepas are in the eastern delta-flank depression of the active Mississippi River delta. Atchafalaya Basin and the many lakes and bays south of it are in the western delta-flank depression. Lake Pontchartrain has subsided in very modern times, as is evident from the Indian village which has been found on its bottom, at a depth of at least 14 feet below existing sea level.

Delta-flank depressions not only indicate easy vertical yielding of the crust but also great horizontal tensile strength.

The absence of delta-flank depressions at the mouths of certain rivers which have changed their courses in recent historic time, for example the Hwang-Ho of China, and Adour of France, suggests that vertical yielding requires a certain minimum of weight in order that it be initiated. Once started the process seems to be readily maintained, as suggested by the fact that bench marks in various parts of the Lower Delta of the Mississippi River subside at rates proportional to proximity of places of heaviest loading (Russell, 1936, p. 175).

The close proximity of the highland of southwestern Mississippi to the site of heaviest sedimentation during modern times is not likely a coincidence. The steep terrace gradients between the most elevated region and the delta-flank depression to the south are unequalled elsewhere in Louisiana (Fisk, 1939a, figs. 3, 5) and suggest an upwarp which is both active and rapid. Drainage pattern changes within very recent times in the territory just west of the uparched structure thoroughly agree with this conclusion (Russell, 1938a, p. 79-86). There seems to be little reason to doubt that the uparching, of which the southwestern Mississippi highland is the summit, has resulted directly from contemporaneous loading of the region of modern deltas.

The similarities in depth, about 3000 feet, between the "bowls" of Recent and Prairie deposition may be more than coincidence. Possibly the strongest evidence against the idea of coincidence comes from the Tertiary section. Fisk has repeatedly pointed out to the writer that 3000 feet is a value approached many times but never exceeded by Tertiary formations in the Gulf Coast. There is thus an isostatic implication that the crust may yield just about this much sedimentary loading and that deposition must then proceed seaward rather than continue thickening of the underlying column. That such an effect is eventually overcome is clear from the record of the Gulf Coast Geosyncline as a whole. The total thickness of the Tertiary is thought to be some 30,000 feet and wells of half that depth are in many cases no deeper, stratigraphically, than the Miocene.

STRUCTURAL IMPLICATIONS

The shape of the Recent formation indicates that subsidence under load may cause rather striking departures from horizontality between various parts of a single formation. Beds laid down horizontally may become so downwarped that dips of over 130 feet per mile, characteristic of parts of the basal gravels of the Recent, occur without the intervention of any subsequent deformation from an outside cause. In other parts of the same formation, as along the more inland parts of the Recent where dips are but 4 inches per mile, horizontality may be maintained.

It seems probable that many lenticular thickenings which have been observed in the formations of exposed geosynclines are examples of the same thing. Initial dips probably have more to do with existing rock attitudes than has been realized.

Localized unconformities in continental sediments may result from such nearby depressions and arches as now exist between the Pontchartrain basin and the southwestern Mississippi highland. There is a great bulk of Pleistocene involved in the uparching and its attitudes and regional patterns are quite different from those of Recent material in the immediate vicinity. Between Pleistocene and Recent formations the stratigraphic relations are both those of disconformity and angular unconformity.

All major characteristics of Cenozoic diastrophic history of the Gulf Coast appear in the Quaternary record. The depositional sequence from fluviatile beds inland to marine facies on the coast involves transition through broad deltaic coastal plain belts, in which dips steepen and formations thicken along well defined zones. Uplift inland has always served to truncate the up-dip portion of formations and subsidence coastward has preserved complete sections.

Briefly stated, the Gulf Coast record is one of continuous sedimentation, southward migrating shore lines, interrupted only slightly by minor advances and retreats of the sea, structural deformation related to sedimentary loading, or in other words a display of the practical workings of uniformitarianism in which the Quaternary is worthy of extremely detailed investigation because of the clarity with which its history may be read.

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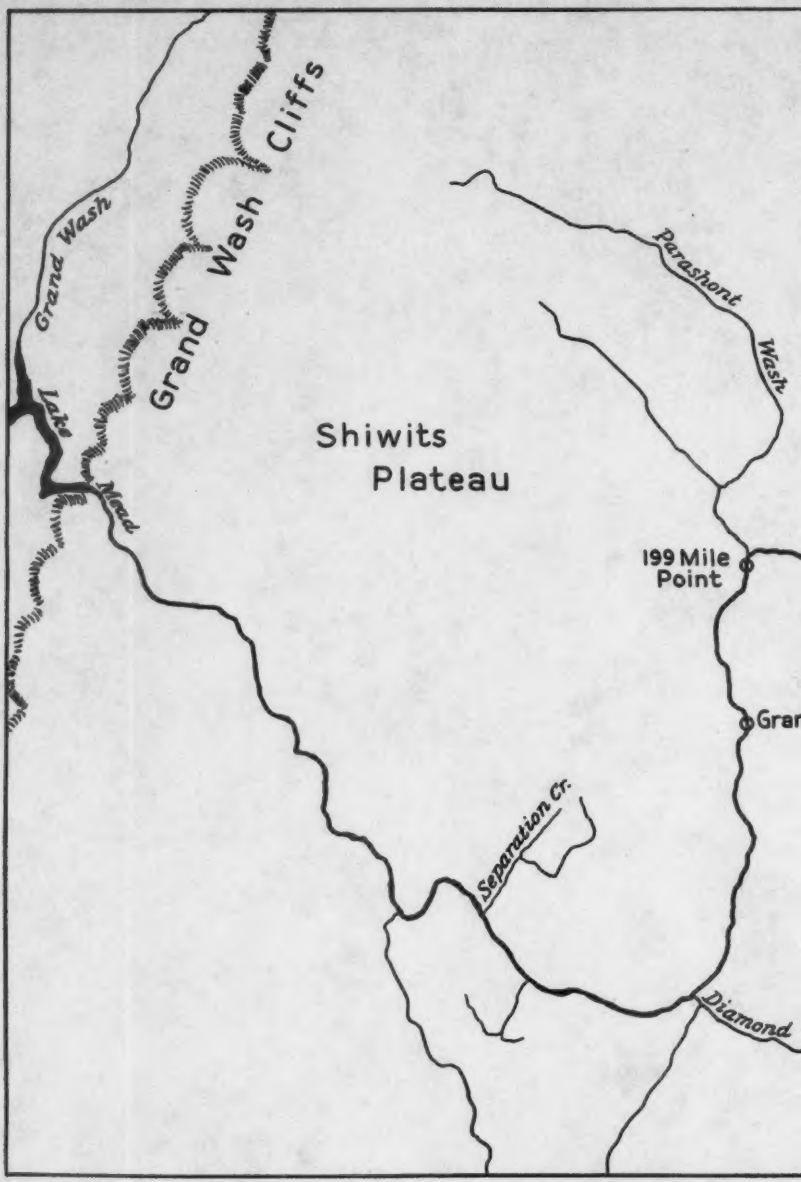
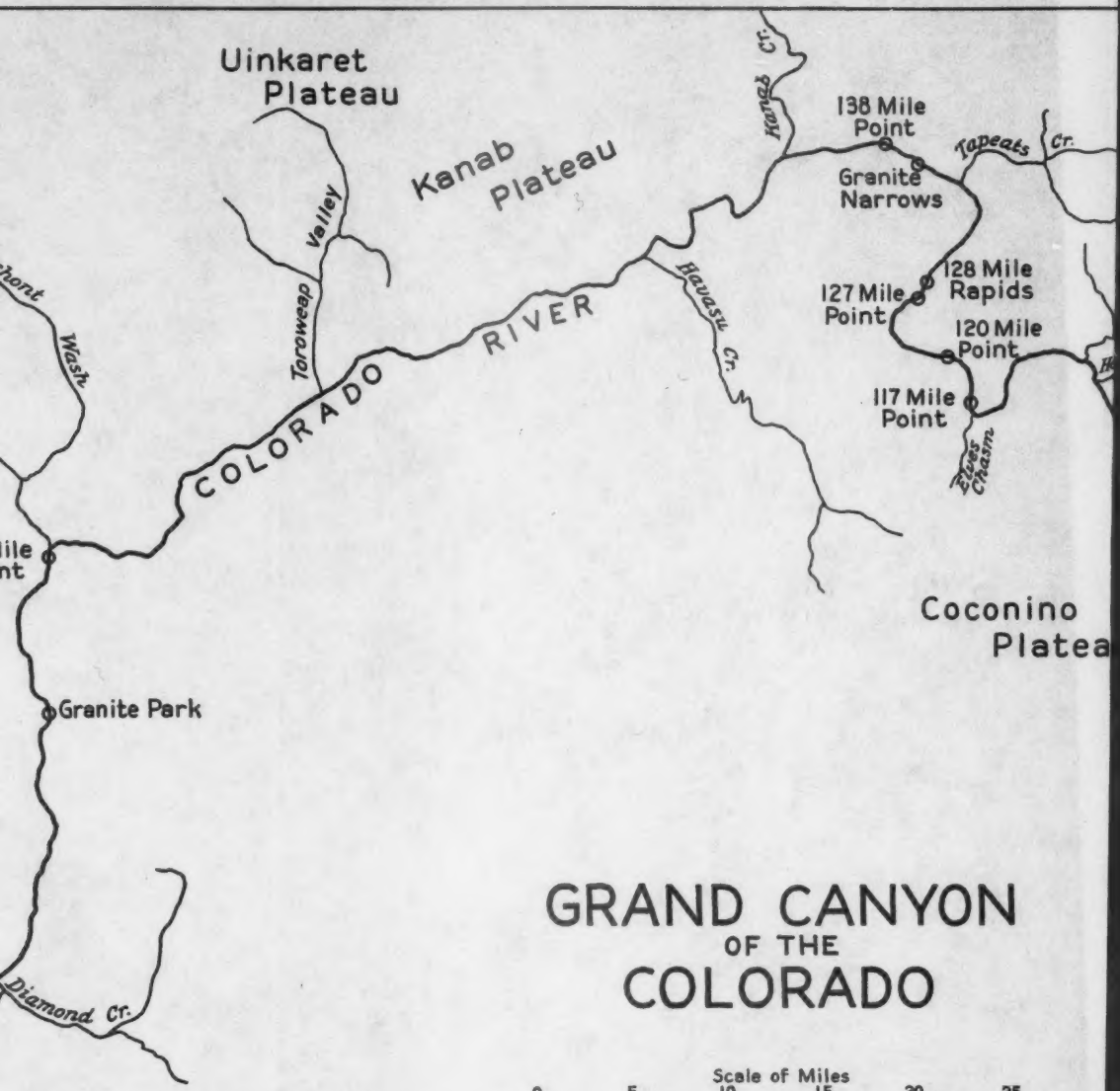
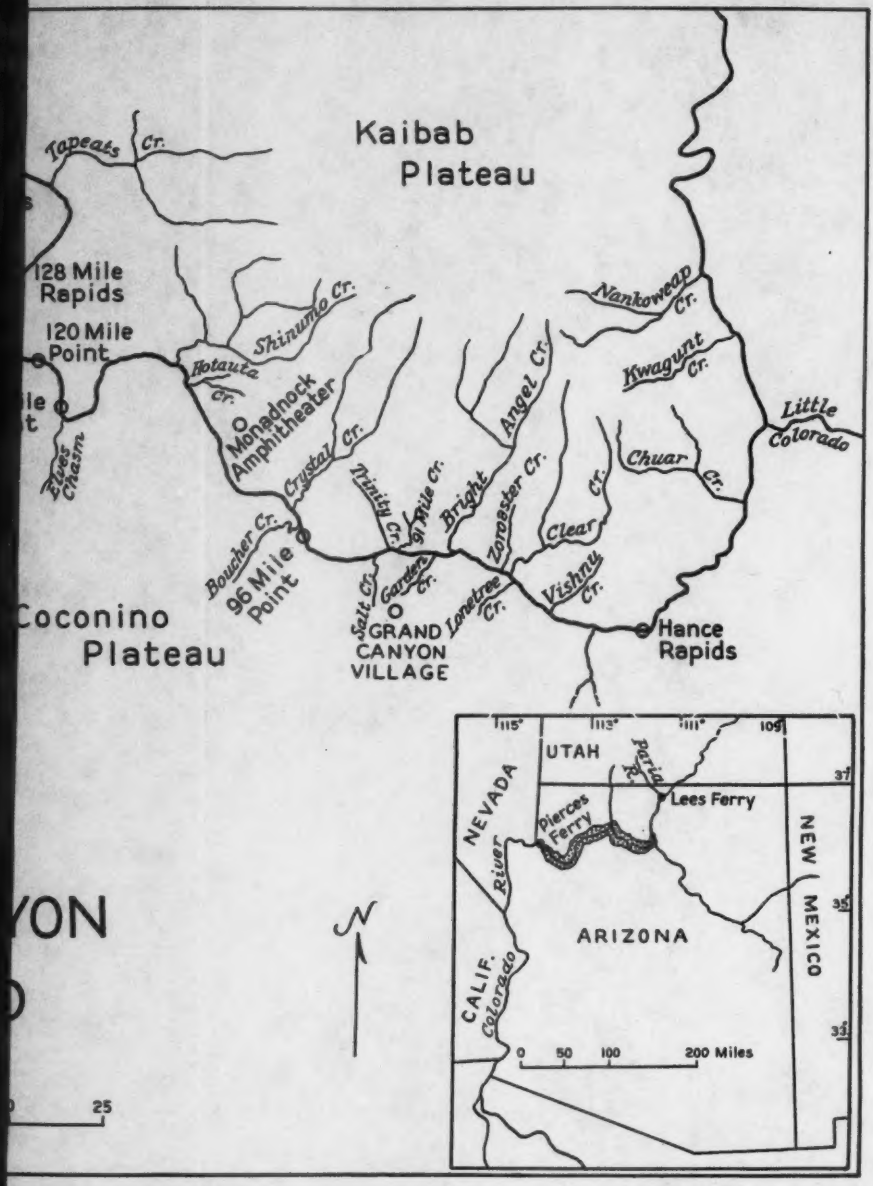

Grand Wash Cliffs
Grand Wash
Lake Mead
Shiwits Plateau
Parashont Wash
199 Mile Point
Separation Cr.
Diamond



MAP OF GRAND CANYON LOCALITIES


Kaibab
Plateau
Tapeats Cr.
128 Mile
Rapids
120 Mile
Point
Shinumo Cr.
Hotauta
Cr.
Monadnock
Amphitheater
Crystal Cr.
Trinity Cr.
3 Mile Cr.
Bright
Angel Cr.
Zoroaster Cr.
Clear
Cr.
Nankoweap
Cr.
Kwagunt
Cr.
Chuar
Cr.
Little
Colorado
River
Chasm
Boucher Cr.
96 Mile
Point
Salt Cr.
Garden
Cr.
GRAND
CANYON
VILLAGE
Lonetree
Cr.
Vishnu
Cr.
Hance
Rapids
Coconino
Plateau
UTAH
NEVADA
ARIZONA
NEW MEXICO
CALIF.
Paria
R.
Lees Ferry
Pierces
Ferry
River
Colorado
0 50 100 200 Miles
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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EP-ARCHEAN AND EP-ALGONKIAN EROSION SURFACES, GRAND CANYON, ARIZONA

BY ROBERT P. SHARP

CONTENTS

	Page
Abstract	1236
Introduction	1236
Introductory statement	1236
Location and physical features	1237
Geological setting	1238
Ep-Archean surface	1239
General statement	1239
Topographic features	1240
Weathering on Ep-Archean surface	1240
Depth and mineralogical changes	1240
Nature of weathering	1242
Hotauta conglomerate	1242
Ep-Algonkian surface	1243
General statement	1243
Topographic features	1244
Weathering on Ep-Algonkian surface	1245
Depth of weathering	1245
Nature of mineralogical changes	1245
Ep-Algonkian regolith	1248
Miscellaneous features	1249
Date of weathering	1250
Nature of weathering	1251
Tapeats sandstone	1251
General statement	1251
Nature of basal Tapeats	1252
Tapeats grit, conglomerate, and breccia	1252
Ep-Archean and Ep-Algonkian climatic conditions	1255
General statement	1255
Evidence on climatic conditions	1255
Red color of weathered material	1255
Feldspars in weathered material	1256
Insoluble residues	1256
Iron oxide content of weathered detritus	1257
Type of weathering	1257
Caliche	1257
Miscellaneous features	1258
Origin of surfaces	1259
Processes and conditions of peneplanation	1259
Thickness of rock removed and duration of erosion intervals	1260

Page

Effects of marine erosion on surfaces 1261
General statement 1261
Modification of Ep-Archean surface 1262
Modification of Ep-Algonkian surface 1262
Summary 1264
Works to which reference is made 1265

ILLUSTRATIONS

Figure Page

1. Diagrammatic cross section, inner part of Grand Canyon 1239
2. Hotauta conglomerate in Hotauta Canyon 1243
3. Ep-Archean and Ep-Algonkian surfaces, Crystal Creek 1246
4. Basal Tapeats and weathered zone on Ep-Algonkian surface, Elves Chasm 1248
5. Residual quartz fragments on Ep-Algonkian surface 1249
6. Weathered zone on Ep-Algonkian surface, Diamond Creek 1250
7. Tapeats breccia and conglomerate lenses on the flank of a monadnock 1254
8. Buried monadnock modified by marine erosion, Hotauta Creek 1263

Plate Facing page

1. Map of Grand Canyon localities 1235
2. Ep-Archean surface 1240
3. Ep-Algonkian surface 1241
4. Ep-Algonkian surface 1244
5. Details of Ep-Algonkian surface 1245
6. Tapeats slide breccia 1254
7. Ep-Algonkian monadnocks 1255

ABSTRACT

The buried Ep-Archean and Ep-Algonkian erosion surfaces of the Grand Canyon are outstanding examples of well-preserved peneplains. The maximum relief of the Ep-Archean surface is less than 50 feet, and monadnocks rise a maximum of 800 feet above the level of the Ep-Algonkian surface, about 95 per cent of which is relatively flat.

Residual regolith on these surfaces shows that they are of subaerial origin and that chemical weathering was dominant in the closing stages of the erosion cycle. The ultimate product of extended weathering consists of quartz, muscovite, clay, and hydrated iron oxides. In some places a residual enrichment of iron appears certain. The most maturely weathered materials contain an average of 88 per cent insoluble residue. Moderately fresh potash feldspar on the surfaces and in the overlying beds is attributed to removal of the more maturely weathered material by the encroaching seas.

Various lines of evidence are cited to show that climatic conditions toward the close of the Ep-Archean and Ep-Algonkian intervals were relatively humid. The most noticeable effects of marine processes are the reworking of the weathered mantle and erosion of monadnocks. A spectacular slide breccia resulting from such erosion is described. A calculation based on the best available evidence indicates approximately 100 million years as the length of the Ep-Algonkian interval, and the Ep-Archean interval may have been longer.

INTRODUCTION

INTRODUCTORY STATEMENT

Geomorphologists have been seriously handicapped in their study of peneplains because such surfaces are usually represented only by scat-

tered remnants of limited extent and so modified by subsequent erosion that the details of form and mode of origin cannot be easily determined. At least a partial solution of this problem appears to lie in the study of buried erosion surfaces. On the steep walls of the Grand Canyon two such buried surfaces are wonderfully exposed. The extent and topographic features of these surfaces are described. The physical character and mineralogical and chemical composition of the overlying and underlying materials have been studied. An attempt is made to present a logical picture of the nature and mode of origin of these surfaces, of the climatic conditions under which they were formed, and of the modifications produced by marine processes when they were submerged.

Following the usage of Lawson (1902, p. 54; 1934, p. 1061-1062) and Hinds (1935, p. 4-7), the names Ep-Archean and Ep-Algonkian will be used. The Ep-Archean erosion surface truncates the Archean complex and is covered by Algonkian strata. The Ep-Algonkian erosion surface truncates both Algonkian and Archean rocks and is buried beneath the Cambrian Tapeats sandstone.

The field study was made during the Carnegie Institution-California Institute of Technology Grand Canyon Expedition of 1937. This expedition travelled by boat down the Colorado River from Lees Ferry, Arizona, to Pierces Ferry, Arizona (inset of Pl. 1). The greater parts of October and November were spent on the river. The principal objective of the trip was a continuation of investigations by Campbell and Maxson (1938a) of the Archean rocks in the Grand Canyon.

Acknowledgment is due the Carnegie Institution of Washington and the California Institute of Technology for sponsorship of the trip. Every member of the expedition, including three boatmen and four geologists, aided this study. Special appreciation is extended to Professors Ian Campbell and J. H. Maxson for opportunity to make this study. Campbell, Maxson, and E. D. McKee have kindly criticised the manuscript, but sole responsibility for the conclusions offered is assumed by the writer. McKee generously offered suggestions and information based on his wide geological knowledge of the region. Mr. D. R. Hanna of the National Youth Administration has assisted in the laboratory. Discussions with the various members of the geology staff at the University of Illinois have proved helpful and are gratefully ackowledged.

LOCATION AND PHYSICAL FEATURES

The Grand Canyon [1] is a mile-deep chasm cut into the plateau area of northwestern Arizona by the Colorado River and its tributaries (Pl. 1).

[1] A proposal before the Geographic Board defines the Grand Canyon as extending from just above Nankoweap Creek westward to the Grand Wash Cliffs, a distance of 224 miles by river. All the features to be discussed are within the boundaries of the Grand Canyon thus defined.

The deepest parts of the Canyon are distinguished by the so-called "inner gorges," which are narrow, steep-walled canyons cut in the Archean complex below the sedimentary rocks. The Ep-Archean and Ep-Algonkian surfaces are exposed near the river or near the upper edges of the inner gorges and in tributary canyons.

Much of the area studied is covered by the Vishnu, Bright Angel, Shinumo, and Supai quadrangles, and the rest, immediately adjacent to the river, is shown on United States Geological Survey profile maps of the Colorado River. Locations on these profile maps are given in terms of the distance from the mouth of the Paria River (Lees Ferry); for example, Bright Angel Creek is at mile 88 and Shinumo Creek at mile 108½. Some of the localities described are named according to this system (Pl. 1).

GEOLOGICAL SETTING

The excellent exposures of Archean, Algonkian, and Paleozoic rocks in the Grand Canyon are widely recognized (Wilson, 1939a, p. 135-136). Only a brief résumé of pertinent features is given.

The Archean complex consists of metasedimentary, metaigneous, and igneous intrusive rocks, including quartzite, sericite quartzite, quartz-mica schist, garnet-sillimanite gneiss, amphibolite, granitized schist, pegmatite, and various plutonic types ranging from granite to tonalite. For further details reference should be made to the works[2] of Noble (1910, p. 381-386), Noble and Hunter (1916), Campbell (1937), and Campbell and Maxson (1933; 1935; 1936; 1937; 1938a; 1938b).

Overlying the Archean with a remarkable unconformity are the strata of the Algonkian[3] Grand Canyon series, preserved in wedge-shaped fault blocks between the Archean and the overlying Cambrian (Fig. 1). The Grand Canyon series comprises almost 12,000 feet of sedimentary rocks with intercalated mafic flows, sills, and dikes. Walcott (1894, p. 506) divided the series into the Unkar and Chuar groups. Van Gundy (1934, p. 345-346; 1937) recognized the Nankoweap group between the Unkar and Chuar. The lower or Unkar group is of significance to this study and has been most completely described by Walcott (1894; 1895), Noble (1910, p. 504-522; 1914, p. 41-60), and Van Gundy (1934).

At the base of the Unkar is the Hotauta conglomerate from a few inches to 30 feet thick. The Hotauta is overlain by the Bass limestone, which is succeeded by the Hakatai shale. The Shinumo quartzite, 1500 feet of massive quartzite and sandstone, overlies the Hakatai. The

[2] The interested reader will find Wilson's (1939a) bibliography of papers on Arizona geology extremely useful.

[3] It seems best in a paper of this type to apply the term Algonkian to the beds of the Grand Canyon series as has been done heretofore, recognizing that the arrangement of pre-Cambrian divisions in the Southwest is in a state of flux (Wilson, 1937; 1939b; Hinds, 1936; 1937a; 1937b; 1937c; 1938a; 1938b; 1938c; Stoyanow, 1936; p. 473-474; 1937, p. 1999-2000) and likely to be revised in the future.

Shinumo is the most resistant formation of the Grand Canyon series, and many of the monadnocks on the Ep-Algonkian surface are underlain by this formation, proving that it was lithified prior to the Ep-Algonkian erosion interval. It is overlain by the Dox sandstone, which in

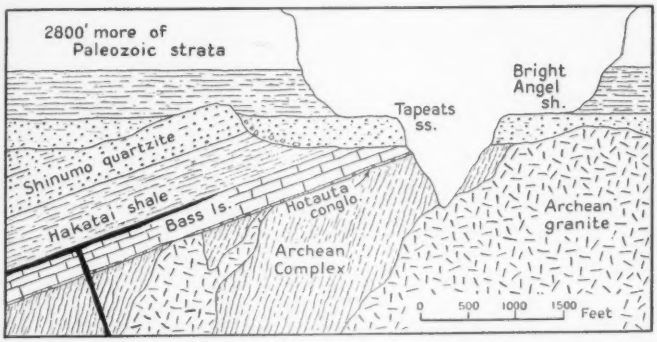

FIGURE 1.—*Diagrammatic cross section, inner part of Grand Canyon*
Hotauta, Bass, Hakatai, and Shinumo are Algonkian; Tapeats and Bright Angel are Cambrian. No direction or elevations given because section is diagrammatic.

turn is succeeded by the Nankoweap and Chuar groups. The Algonkian strata have been deformed by faulting and broad folding (the Grand Canyon disturbance) and extensively eroded prior to deposition of the Cambrian Tapeats sandstone, basal member of the Paleozoic beds.

The Tapeats, a quartzitic sandstone, rests unconformably on both Archean and Algonkian rocks (Fig. 1). It has a maximum thickness of 325 feet and gradually thins westward with local variations owing to irregularities of the surface upon which it was deposited. The Tapeats is overlain by the Bright Angel shale with a maximum thickness of 400 feet, although it is much thinner over the crests of the higher Ep-Algonkian monadnocks. Detailed descriptions of the Tapeats and Bright Angel have been given by Noble (1914, p. 61-64; 1922, p. 37-42), Schuchert (1918), and Wheeler and Kerr (1936, p. 6-15). The remainder of the Paleozoic column in the Grand Canyon is not treated because it is of no particular significance to this study. Figure 1 shows in diagrammatic fashion the geologic relations in the inner part of the Grand Canyon.

EP-ARCHEAN SURFACE

GENERAL STATEMENT

The remarkably smooth erosion surface developed on the Archean complex and buried beneath Algonkian beds is the Ep-Archean surface.

It was first described by Powell (1875, p. 212-213) and Walcott (1890, p. 57; 1894, p. 507, 512; 1895, p. 317, 318, 324) and has been mentioned by many subsequent writers. Noble (1910, p. 504, 523-524; 1914, p. 41-42, 80) and Hinds (1935, p. 8-21) have given the most detailed descriptions of this surface.

The Ep-Archean surface is preserved only where blocks of Algonkian strata, faulted down into the Archean, have escaped removal during the Ep-Algonkian erosion interval. For this reason localities where it can be observed are few and widely separated. However, the excellence of exposure and the spacing of localities along a traverse distance of 57 miles give a fairly complete picture. The topography of the surface, the weathering of the underlying Archean rocks, and the nature of the overlying beds will be described briefly.

TOPOGRAPHIC FEATURES

The extreme smoothness of this surface is its principal feature. Noble (1910, p. 504; 1914, p. 42) notes a maximum relief of 20 feet in the Shinumo quadrangle, and Hinds (1935, p. 9) reports a relief of less than 50 feet in the area investigated by his parties, chiefly the Bright Angel and Vishnu quadrangles. Below Hance Rapids (Pl. 2, fig. 1), on Crystal Creek (Fig. 3), in Hotauta Canyon, and at Granite Narrows (Pl. 2, fig. 2) the Ep-Archean surface is extremely smooth with a relief of only a few feet. Opposite the mouth of Shinumo Creek low rounded knobs of Archean rock project 20 feet above the general level, and this was the maximum relief observed.

The extremely low relief of this surface may mean that it approaches the ultimate form of the erosion cycle. This ultimate form—the plain—has probably never been recognized except in theory (Davis, 1899b, p. 497); therefore, the Ep-Archean surface is usually referred to as a peneplain. Its mode of origin is considered in a later section.

WEATHERING ON EP-ARCHEAN SURFACE

Depth and mineralogical changes.—The Archean rocks are noticeably weathered to an average depth of 10 feet below the Ep-Archean surface, although locally at the foot of Hance Rapids they are relatively fresh just below the surface. This relation is thought to be due to removal of weathered detritus by the Algonkian sea.

A suite of specimens collected from the Archean at the head of Granite Narrows (Pl. 2, fig. 2) illustrates the mineralogical changes in the weathered zone. At this point the Archean consists of mica schist, granitized schist, micaceous quartzite, pegmatite, and granite. A reasonably fresh sample of muscovite-biotite granite containing moderately large crystals of microcline has been obtained 10 feet below the uncon-

FIGURE 1. EP-ARCHEAN SURFACE BELOW HANCE RAPIDS
Ep-Archean surface (dashed line) between Archean complex and overlying, tilted Algonkian beds. View eastward upstream.

FIGURE 2. EP-ARCHEAN SURFACE AT GRANITE NARROWS
Ep-Archean surface (dashed line) and Ep-Algonkian surface (dotted line). View westward downstream.

FIGURE 1. EP-ALGONKIAN SURFACE IN EASTERN PART OF GRAND CANYON
Horizontal Paleozoic strata separated from tilted Algonkian beds by Ep-Algonkian surface (dashed line). View southeast. (Photograph by courtesy of U. S. Department of Interior.)

FIGURE 2. EP-ALGONKIAN SURFACE BELOW 137 MILE RAPIDS
Horizontal Tapeats sandstone beds resting upon featureless Ep-Algonkian surface developed on Archean complex.

EP-ALGONKIAN SURFACE

formity. The only alteration in this rock is minor bleaching and chloritization of the biotite, and some sericite and kaolin on the plagioclase. This same rock 5 feet below the unconformity is loose and crumbly and shows evidence of considerable weathering. Nearly all the biotite is completely chloritized, and the plagioclase has been almost entirely decomposed, although the potash feldspar is relatively bright and shows only minor alteration. The rock is stained a light brown by iron oxide.

Unfortunately, this lithologic type does not extend to the surface of the unconformity, but at this same locality a quartz-mica schist 3 feet below the unconformity shows alteration of biotite to chlorite, and the whole mass is stained a dark reddish brown by iron oxides which constitute 3.5 per cent of the rock. This same rock at 1½ feet below the Ep-Archean surface consists chiefly of quartz and muscovite, a minor amount of clay and chlorite, and 2.5 per cent iron oxides. Six inches below the surface the weathered material consists of quartz, muscovite, clay, and 2.5 per cent iron oxides; chlorite is destroyed. The higher percentage of iron oxides at 3 feet may indicate leaching near the surface and deposition at depth. Other samples of the weathered material below the Ep-Archean surface show essentially the features just outlined.

The destruction of plagioclase feldspar and of iron containing minerals, chiefly biotite, and an enrichment in residual quartz and muscovite is notable in the weathered zone. Potash feldspar is decomposed and decreases in amount so that in some places it is entirely eliminated. That only a minor part of this weathering can be attributed to recent agents is demonstrated in the consideration of weathering on the Ep-Algonkian surface where more data are at hand.

Patches of residual regolith [4] developed by subaerial weathering remain on the Ep-Archean surface where they have not been removed by the Algonkian sea. Such residual regoliths are preserved in Hotauta Canyon and at Granite Narrows. The typical regolith is a dark-reddish, ferruginous layer usually a few inches to a foot thick in which the structures of the underlying rocks are no longer apparent. It consists

[4] Merrill (1897, p. 299) proposed the term regolith for the entire mantle of unconsolidated material on the earth's surface, whatever its nature and origin. He definitely included both residual and transported material in his definition. The term has been used in some textbooks solely for the residual mantle rock (Pirsson and Schuchert, 1924, p. 196; Chamberlin and Salisbury, 1909, p. 472; Worcester, 1939, p. 131). However, other writers, for example Leith and Mead (1915, p. xxi), Tarr and Martin (1918, p. 30), Ries and Watson (1930, p. 214), and Tolman (1937, p. 127) use the term in the sense originally proposed by Merrill.

Becker (1895, p. 289) suggested the term saprolite for strictly residual mantle rock. Merrill (1897, p. 301) objected to this term because the Greek root means rotten and refers to the decomposition of organic material. Sederholm (1931, p. 77) suggested the term sathrolite, from the Greek root meaning weathered, for strictly residual weathered mantle rock. Sederholm's term is probably the best of the group, but it is not widely used in the United States. For this reason the term regolith is used in this report, but, in order to define the term more sharply, adjectives such as residual and transported will be added. If the exact nature of the weathered mantle is unknown or if it is both residual and transported, the term regolith will be used without modifiers.

in major part of minerals resistant to weathering, usually quartz and muscovite, and may contain some potash feldspar, clay, and considerable secondary iron oxide.

Nature of weathering.—In addition to the chemical changes just discussed the weathered detritus is somewhat loose and crumbly, a condition which could be attributed to granular disintegration, a mechanical process. However, chemical weathering, particularly hydration, can make rocks crumbly, and as shown by Blackwelder (1925; 1927) the so-called mechanical processes of weathering may actually be partly chemical. The evidence of chemical weathering on the Ep-Archean surface appears to be far greater than any reliable evidence of mechanical weathering.

HOTAUTA CONGLOMERATE

The Hotauta conglomerate [5] in most places is of local origin, consisting of angular to subangular fragments derived from underlying or near-by Archean rocks. The Hotauta is a dark reddish-brown, poorly bedded and poorly sorted sandstone or conglomerate, locally arkosic, and of poor to strong coherence. It has a maximum thickness of 30 feet, although the average is between 2 and 6 feet, and in some places it appears to be absent.

In Hotauta Canyon at the locality from which the formation was named by Noble (1914, p. 42) the Hotauta is 2 feet thick and consists chiefly of reworked arkosic detritus derived from underlying weathered Archean quartz diorite (Fig. 2). Within the reworked material are fragments of quartz diorite several inches in diameter. The matrix of the basal Hotauta contains particles showing some rounding and wear. It is calcareous, a fact probably to be attributed to downward leaching and deposition of calcite from the overlying Bass limestone, and contains less than 1 per cent iron oxides, a surprisingly small amount in view of its intense reddish-brown color.

On Crystal Creek the Hotauta is a few feet thick and consists of a coarse breccia at the base and a red sandstone in the upper part. The basal breccia is composed of angular fragments derived from the underlying sericite-biotite schist and quartzite. At Granite Narrows the Hotauta is a 2-foot bed of reddish-brown quartzite containing some muscovite and 2 per cent iron oxides. This bed has been derived largely from the underlying residual Archean regolith which consists of quartz, muscovite, clay, and iron oxides.

[5] Van Gundy (1934, p. 342) and Hinds (1936, p. 106-107) include the basal Algonkian conglomerate and sandstone in the Bass limestone and discard Hotauta as a formational name. The use of the name Hotauta is continued here for the sake of convenience without necessarily accepting the unit as a formation.

The thickest exposure of Hotauta noted in this study is at the foot of Hance Rapids. At this locality it is a dark-brown, coarse conglomerate about 20 feet thick consisting of subangular to rounded pebbles, cobbles, and boulders up to 1½ feet in diameter of pegmatite, granite, aplite, rhyolite, quartz, quartzite, and quartz-mica schist, cited in order of

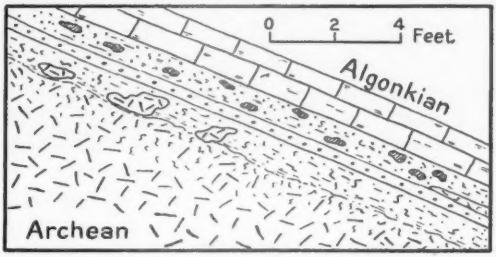

FIGURE 2.—*Hotauta conglomerate in Hotauta Canyon*
Hotauta is overlain by the Bass limestone; residual boulders at base of Hotauta.

abundance, in a well-cemented matrix of coarse arkosic sand. The base of the conglomerate consists of very angular fragments of the underlying Archean. The quartzite composing the pebbles of this conglomerate is notably different from the quartzite of the near-by Archean, and no rock type even remotely resembling the rhyolite pebbles is known in the Grand Canyon Archean terrane. Hinds (1935, p. 23; 1936, p. 94, 99, 103) suggested that the quartzite and rhyolite in the Hotauta have been derived from the Mazatzal Mountains area of central Arizona, 120 miles to the south, where such rock types, presumably older than the Grand Canyon series (Wilson, 1939b, p. 1120, 1124-1127), are known.

The Hotauta is overlain conformably and gradationally by the Bass limestone, and in some places the Bass appears to rest directly upon the Ep-Archean surface. The evidence of reworking of the Ep-Archean regolith and the presumably marine origin of the Bass limestone, although perhaps not entirely conclusive, suggest that the Hotauta is the initial deposit of the Algonkian sea which swept across the Ep-Archean peneplain (Barton, 1916, p. 446; Noble, 1914, p. 80-81).

EP-ALGONKIAN SURFACE

GENERAL STATEMENT

The Ep-Algonkian surface is easily observed by Grand Canyon visitors, for the unconformable contact between the horizontal Tapeats sandstone

and the pre-Cambrian rocks is visible from the edges of the Coconino or Kaibab plateaus (Pl. 3, fig. 1). It was early described by Newberry (1861, p. 57-58), Powell (1875, p. 212), Gilbert (1875, p. 51, 54-55), Dutton (1882, p. 179, 181, 209), and Walcott (1883, p. 439; 1890, p. 54-58) and has been the subject of comment by many subsequent writers. Noble (1910, p. 504, 526-527; 1914, p. 40, 52, 61-62, 83-84; 1922, p. 38), Hinds (1935), and Wheeler and Kerr (1936, p. 2-5) have given the most complete description. In this study the Ep-Algonkian surface has been examined at localities along a traverse distance of 210 miles which is included in a roughly rectangular area, 100 miles east-west and 50 miles north-south. The exposures, although not continuous throughout this distance, were excellent, and exposures continuous for many miles were observed.

The Ep-Algonkian surface has considerably greater topographic relief than the Ep-Archean surface. It truncates both Archean and Algonkian rocks, but at least 95 per cent of the underlying rocks are Archean. In places a residual regolith is preserved on the surface; elsewhere most of the regolith has been incorporated in the basal Tapeats sandstone.

TOPOGRAPHIC FEATURES

The relief of the Ep-Algonkian surface is controlled by the nature of the underlying rocks, for the highest monadnocks are composed of Shinumo quartzite (Pl. 7, fig. 1) and massive granite. The monadnocks underlain by the Shinumo quartzite are linear strike ridges (Noble, 1914, p. 40; McKee, 1933, p. 203-204; Hinds, 1935, p. 27; Wheeler and Kerr, 1936, p. 4-5) which form low hogbacks or cuestas on the Ep-Algonkian surface. The monadnocks of Archean rocks, chiefly massive granite, are lower, more rounded, and of no great lateral extent, as exemplified by a monadnock just west of Zoroaster Creek in the Bright Angel quadrangle (Pl. 4, fig. 1).

Hinds (1935, p. 27) reports a maximum relief of 250 feet in the Bright Angel and Vishnu quadrangles, and Noble (1914, p. 51) records a relief of 600 feet in Monadnock Amphitheater in the Shinumo quadrangle. A large monadnock on the north side of the Grand Canyon between Bright Angel and Ninetyone Mile creeks (Pl. 6, fig. 2) is almost 800 feet high (Wheeler and Kerr, 1936, p. 5), and this is the highest known point on the Ep-Algonkian surface in the Grand Canyon, where more than 95 per cent of the Ep-Algonkian surface has a relief of less than 150 feet and a large part is essentially flat (Pl. 3, fig. 2). Some students may wish to call such a surface an "old land" (Maxson and Anderson, 1935, p. 90-91), and others may prefer to call it a peneplain. In this paper the Ep-Algonkian surface will be referred to as a peneplain with the realization that not everyone will agree.[6]

FIGURE 1. EP-ALGONKIAN SURFACE NEAR ZOROASTER CREEK
Tapeats sandstone wedges out both to the right and left. Monadnock of massive Archean granite on right.

FIGURE 2. TAPEATS-ARCHEAN CONTACT, NINETYONE MILE CREEK
Fine-grained basal Tapeats beds resting on Archean granite and overlain by coarse slide breccia; *see also* Plate 6.

EP-ALGONKIAN SURFACE

FIGURE 1. BOULDERS OF DECOMPOSITION IN ARCHEAN GRANITE BELOW EP-ALGONKIAN SURFACE
Dark projection at top of photograph is Tapeats sandstone. 96 mile point.

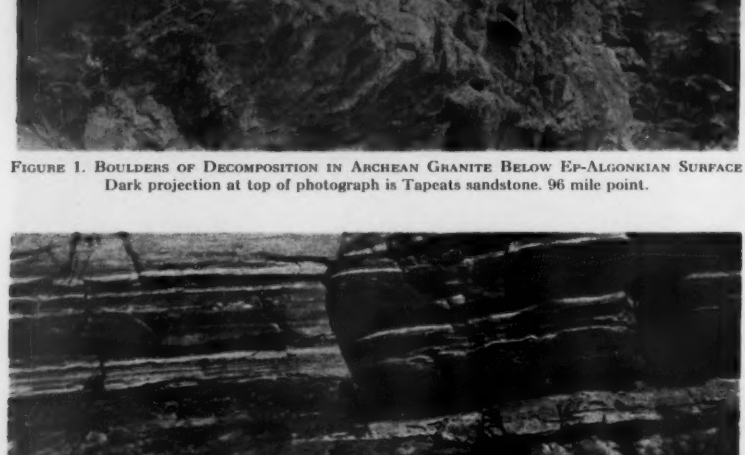

FIGURE 2. BASAL TAPEATS SANDSTONE ON WEATHERED ARCHEAN COMPLEX
Garden Creek Indian ruins.

DETAILS OF EP-ALGONKIAN SURFACE

WEATHERING ON EP-ALGONKIAN SURFACE

Depth of weathering.—Extensive weathering of Archean rocks, chiefly gneissic granite 10 to 12 feet below the Ep-Algonkian surface, has been observed in many places, notably at the Garden Creek Indian ruins (Pl. 5, fig. 2), in Ninetyone Mile Canyon, in Elves Chasm (Fig. 4), and elsewhere. At the 96 and 117½ mile points Archean granite shows minor effects of weathering 20 to 25 feet below the surface. Archean mica schist and granitized schist are oxidized and stained to a depth of 50 feet in a small canyon opposite the mouth of Vishnu Creek, in Lonetree Canyon, and at mile 128. In the first small canyon east of Salt Creek Archean amphibolites and quartz-biotite schists are weathered 40 feet below the unconformity. On Crystal Creek the Tapeats sandstone rests on Algonkian beds (Fig. 3). The rocks below the contact show little evidence of weathering, and this appears to be the case in other localities where the Tapeats-Algonkian contact has been observed. The Algonkian rocks show only slight weathering, probably because they are composed of materials which have already survived at least one cycle of weathering.

In brief, the rocks below the Ep-Algonkian surface show extensive weathering to a depth of 10 to 12 feet and in some places are noticeably weathered 50 feet below the surface. The original thickness of the weathered layer was somewhat greater owing to the removal of material by the Tapeats sea. Algonkian rocks show less weathering than the Archean rocks, and among the Archean rocks mica schist shows weathering to the greatest depth.

Nature of mineralogical changes.—The nature of the mineralogical changes attributable to weathering on the Ep-Algonkian surface can best be illustrated by the following description.

In Ninetyone Mile Canyon (sheet C, Colorado River profile maps) the Archean consists of a fine-grained gneissic granite, granitized schist, and

[6] It is not the purpose of this paper to discuss just what may be called a peneplain. The propriety of referring to the Ep-Algonkian surface as a peneplain has been questioned by McKee (Wheeler and Kerr, 1936, p. 4) and Wheeler and Kerr (1936, p. 4-5). This objection is based on the fact that in at least the eastern part of the Grand Canyon the number and height of the residual ridges on the Ep-Algonkian surface is too great to permit it to be a peneplain, and furthermore that these monadnocks were probably considerably higher at the beginning of the Cambrian submergence. The reason for this being that the higher monandocks were not completely submerged until after all of the Tapeats and a considerable part of the Bright Angel had been deposited, and therefore they were reduced by subaerial erosion throughout a period presumably extending from far down in the Lower Cambrian well into the Middle Cambrian. In his original discussion Davis (1899a) did not give definite limits to the relief permissible on a peneplain. However, in a later paper he (Davis, 1922, p. 588) did treat this matter in a general way by discussing the topographic nature and general elevation of a peneplain, and from this discussion it might be concluded that the Ep-Algonkian surface could be called a peneplain. In support of this view it can be noted that Davis (1901, p. 174-175; 1909a, p. 369; 1913, p. 313) referred to the Ep-Algonkian surface as a peneplain. He recognized the monadnocks on this surface, although he may not have been aware of their number or extent. Maxson and Anderson (1935, p. 90-91) recognized the need for more precise definition of surfaces in the late stage of the erosion cycle and suggested "old land" for a surface in a late stage of the erosion cycle when a relief of several hundred feet may exist.

quartz-mica schist. The rocks are notably weathered to a depth of 10 feet. The biotite in the granite at 10 feet is almost entirely chloritized, and the feldspars show some kaolinization. At 5 feet below the unconformity the biotite is replaced by masses of relatively clear chloritelike

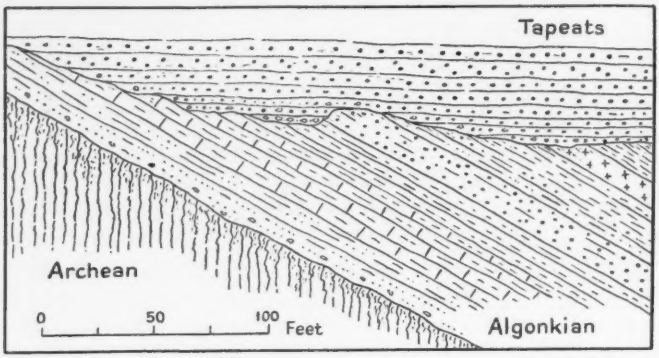

FIGURE 3.—*Ep-Archean and Ep-Algonkian surfaces, Crystal Creek*
Comparative smoothness of the two surfaces notable.

material containing numerous dispersed grains of hematite. Grains of iron oxide are scattered throughout the rock, and the feldspars show some alteration and decrease in amount. At 3 feet below the unconformity the biotite, chlorite, and feldspars have disappeared, and the rock consists of quartz, muscovite, and iron oxides. At 3 inches the rock is essentially the same as at 3 feet, except that at 3 inches it is richer in iron oxides and contains 84 per cent insoluble residue. The percentages of iron oxides in the samples are as follows: at 10 feet, 2.5; at 5 feet, 4.3; at 3 feet, 4.6; at 3 inches, 12.8.

Weathered material on the Ep-Algonkian surface has been studied in many localities of which some of the best are: opposite the mouth of Vishnu Creek, Garden Creek Indian ruins, at mile 96, in Elves Chasm, near Salt Creek, at 128 Mile Rapids, at mile 117½, in Granite Park, on Diamond Creek (Fig. 6), and in Separation Canyon. In all places development of green chlorite from biotite is widespread, and in some places this chlorite is filled with needles of rutile, suggesting that the original biotite was titaniferous. In the more intensely weathered rocks green chlorite is replaced by a clear chloritelike mass containing dispersed grains of iron oxide (probably hematite) which originally may have been magnetite. Other iron-bearing minerals such as hornblende have weathered rather readily. Plagioclase feldspars are usually eliminated,

and at some localities even the potash feldspars have been entirely destroyed, although elsewhere they show only a moderate degree of alteration.

Table 1 gives the percentages of iron oxides [7] in various samples. Camp-

TABLE 1.—*Percentages of iron oxides in weathered detritus*

Type of material	Location	Percentage of iron oxides
Residual regolith on Archean schist and granitized schist	Opposite mouth of Vishnu Creek	4.10
Weathered Archean gneissic granite and granitized schist, 1 foot below unconformity	Garden Creek trail	3.30
Weathered gneissic Archean granite, 6 inches below unconformity	Garden Creek Indian ruins	3.70
Weathered Archean granite	Ninetyone Mile Canyon	
10 feet below unconformity		2.50
5 feet below unconformity		4.30
3 feet below unconformity		4.60
3 inches below unconformity		12.80
Weathered Archean gneiss, 10 feet below unconformity	Mile 117½	7.20
Archean schist, 6 inches below unconformity	127 Mile Rapids	7.80
Same, 30 feet below unconformity	127 Mile Rapids	5.57

bell and Maxson (1935, p. 324) have published some chemical analyses of Archean rock types in the Grand Canyon. The granite analyses contain 1.61 to 1.98 per cent FeO and Fe_2O_3, and one granodiorite contains 6.96 per cent FeO and Fe_2O_3. The metamorphic rocks contain 1.92 to 15.63 per cent FeO and Fe_2O_3, and one ferruginous quartzite contains the unusually high amount of 37 percent, chiefly Fe_2O_3. It is clear from

[7] The method of Zimmermann and Rheinhardt as outlined by Treadwell and Hall (1930, p. 515-520) was used to determine the iron. This procedure involves titration of a solution of ferrous chloride with potassium permanganate in the presence of manganous sulphate. The iron oxides in the weathered material are probably of several varieties including hematite, limonite, goethite, and turgite. The various types have not been separated, and all the iron has been calculated as Fe_2O_3. Undoubtedly, a good part of the oxides are hydrous, and the addition of water to the calculations would add as much as 15 to 20 per cent to the figures in Table 1. Tests were made on biotite and magnetite to determine how unweathered iron containing minerals might affect the calculations, and it was found that these minerals were not soluble enough to produce any great difference.

these analyses that the percentages of iron oxides cited in Table 1 may or may not indicate a residual concentration depending upon the nature of the original material. The data from Ninetyone Mile Canyon, where the amount of iron oxide in the weathered material increases progressively

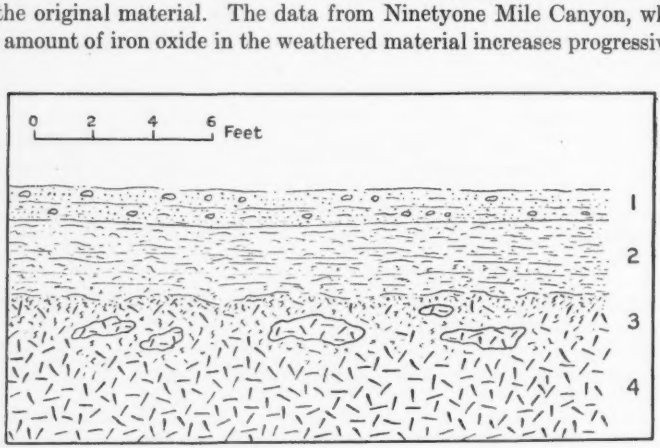

FIGURE 4.—*Basal Tapeats and weathered zone on Ep-Algonkian surface, Elves Chasm*

(1) Gritty Tapeats sandstone. (2) Reworked weathered detritus. (3) Weathered zone with boulders of decomposition. (4) Fresh Archean granite.

upward toward the erosion surface and reaches a maximum of 12.8 per cent, indicate a residual enrichment, for the unweathered gneissic granite at Ninetyone Mile Canyon probably contains considerably less than 12.8 per cent iron calculated as the oxide. Other cases of residual enrichment in iron oxides are suspected but not definitely proved.

At mile 96 the upper 2 or 3 feet of the weathered layer is a light green which contrasts with the dark reddish brown of the weathered rock just below. Presumably the iron oxides have been leached from this uppermost zone, the green color being due to epidote. A light-colored band was noted on the Ep-Algonkian surface just west of Zoroaster Creek. This too may be a leached zone, but it was not examined at close range.

Ep-Algonkian regolith.—In places the Ep-Algonkian surface is mantled by a foot or two of structureless, extensively weathered detritus which passes gradationally upward into the overlying Tapeats and downward into less weathered Archean rock. This layer is considered to be a residual regolith formed by subaerial weathering on the Ep-Algonkian surface. In Elves Chasm the boundary between a well-developed residual regolith and the overlying Tapeats sandstone is determined with considerable difficulty. The presence of worn and somewhat rounded quartz

grains is believed to indicate reworking, and the reworked material is probably Tapeats. This regolith grades downward into less weathered rock which contains cores of sounder material resembling boulders of decomposition (Fig. 4). Similar residual regoliths have been observed

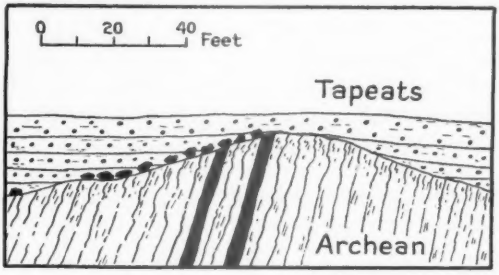

FIGURE 5.—*Residual quartz fragments on Ep-Algonkian surface*

Near 120 Mile Rapids. Solid black represents quartz veins and fragments. Scale applies only to height of hill.

opposite Vishnu Creek, in Ninetyone Mile Canyon, and at other places. They are composed of the more resistant minerals in the underlying rocks and are rich in iron oxides. The original thickness of the regolith cannot be determined, owing to erosion by the Tapeats sea, but may have amounted to several feet in favored localities.

According to Byers, Kellogg, Anderson, and Thorpe (1938, p. 961) normal or zonal soils are those that owe their most important characteristics to the effects of the climate and biological factors acting upon them. Intrazonal soils owe their distinguishing features to the overbalancing effects of parent material or relief. Azonal soils are those that have few or no soil morphological characteristics. Their characteristics are similar to those of the parent material. Most of the Ep-Archean and Ep-Algonkian weathered material is probably intrazonal or azonal, but some of the most maturely weathered detritus may be normal soil.

Miscellaneous features.—On the southwest wall of the Grand Canyon at mile 96 large rounded boulders of decomposition have been formed in the Archean granite beneath the Ep-Algonkian surface. As shown in Figure 1 of Plate 5, the boulders are limited to a zone approximately 10 feet thick below the Tapeats. Weathering, probably largely chemical, along fractures and joints has left cores of relatively unaltered rock which are truly boulders of decomposition. Boulders of decomposition have

also been observed elsewhere but are not so well developed as at this locality.

At mile 119 surficial creep has reversed the dip of the schistosity in weathered schist just below the Tapeats on the flanks of a monadnock.

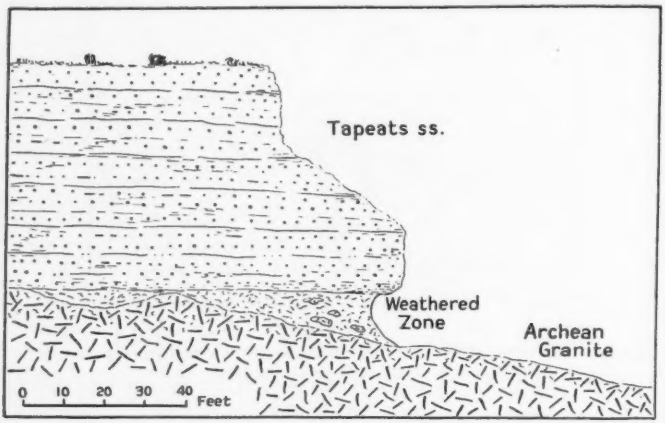

FIGURE 6.—*Weathered zone on Ep-Algonkian surface, Diamond Creek*
Weathered zone is less resistant to modern weathering than surrounding rocks.

Judging from the relations at this place and at mile 127½ the movement is evidently pre-Cambrian and resembles similar creep on modern slopes.

Below 120 Mile Rapids a low rounded knob of Archean schist projects into the basal Tapeats. Near the crest of the knob two quartz veins crop out, and angular fragments of quartz derived from these veins now lie on the slopes of the knob (Fig. 5). If the sandstone were removed, the relations would be essentially those on any present-day hillock containing an exposed quartz vein.

The weathered zone at the top of the Archean is so much softer than the overlying Tapeats and the underlying unweathered rock that caves and alcoves are commonly developed by recent erosion along it. Excellent examples of such caves have been noted in Elves Chasm and at mile 117. The topographic effect of the weathered zone on Diamond Creek is shown in Figure 6.

Date of weathering.—The possibility that the weathering beneath the Ep-Archean and Ep-Algonkian surfaces is largely recent or subrecent is dismissed for the following reasons: (1) In many places the weathered material on the surfaces is a residual regolith. (2) This regolith has been

reworked and incorporated in the overlying sedimentary deposits, hence it must have been in existence prior to encroachment of the seas in which those deposits were formed. (3) The beds overlying the Ep-Archean surface contain fragments of gossanlike material derived from the underlying weathered zone. (4) Residual quartz fragments on hillocks and surficial creep indicate ancient weathering on at least the Ep-Algonkian surface. (5) Recent weathering has developed caves and alcoves along the weathered zone, it has not produced that zone.

Some of the chloritization of biotite and the kaolinization of feldspar may be attributed to recent weathering. However, the progressive weathering of biotite, already described, and its relations to the erosion surfaces indicate ancient weathering. Furthermore, the feldspar in the basal Tapeats is usually much fresher than in the underlying Archean from which it has been derived, showing that the kaolinization of the feldspar cannot be attributed entirely to recent weathering. Presumably, the feldspar in the Tapeats has been freshened by removal of altered material during the reworking.

Nature of weathering.—Within the Ep-Algonkian weathered zone the development of iron oxides, the elimination of feldspar, biotite, and other iron-bearing minerals, and the high percentage of insoluble residues indicate considerable chemical weathering. Furthermore, the boulders of decomposition are chiefly the product of chemical weathering. Even the crumbly nature of the weathered rock may be in large part due to disruptions caused by changes of volume as minerals were altered. Mechanical processes of weathering have probably acted on the Ep-Algonkian surface especially on the sides of monadnocks, but their effects are subordinate. It must be remembered, however, that the effects of only those processes active in the final stages of the cycle are preserved. Early in the cycle mechanical weathering may have been dominant.

TAPEATS SANDSTONE

General statement.—The Tapeats sandstone is an irregularly bedded, cliff-forming unit. It is a tan, white, reddish, or brown sandstone or grit, locally arkosic, which in places is cemented well enough with silica to qualify as a quartzite. The Tapeats has a maximum thickness of 325 feet, although locally, where Ep-Algonkian monadnocks rose above its level of deposition, the formation is missing (Pl. 7, figs. 1, 2). Irregular stratification, cross-bedding, and coarse grain size are in accord with the belief that it was deposited in shallow water. It has long been considered Middle Cambrian (Noble, 1922, p. 39) on the basis of fossils supposedly collected from its upper part; however, Wheeler and Kerr (1936, p. 12-14) and McKee (1937, p. 341; 1938, p. 358) report that the Tapeats in at

least the western and central parts of the Grand Canyon is Lower Cambrian and that it becomes progressively younger eastward owing to overlap, so that the actual time involved in deposition of the Tapeats probably includes a large part of the Lower Cambrian.

Nature of basal Tapeats.—The basal part of the Tapeats is of major interest because of its relation to the Ep-Algonkian surface. In many localities the basal Tapeats is composed almost entirely of reworked regolith, of which particularly good examples were observed at the Garden Creek Indian ruins (Pl. 5, fig. 2) and in Elves Chasm (Fig. 4). At both places the reworked material is distinguished from the underlying weathered rock only by a crude bedding, by faint rounding or wear of some of the particles, or by a small amount of material introduced from near-by outcrops. Usually only the basal 2 or 3 feet of the Tapeats is dominantly reworked regolith, and this basal zone is succeeded by alternating dark and light bands. The dark bands consist of ferruginous weathered material reworked from near-by outcrops, and the light bands are fresh material derived from an area which had already been cleaned of its weathered material. The weathered detritus is usually limited to pocketlike deposits in the lower 10 feet of the Tapeats, although weathered material has occasionally been observed 50 feet above the base. The reworked material has about the same mineralogical composition as the underlying regolith and usually contains a smaller amount of iron oxides as illustrated in Table 2. The Garden Creek Indian ruins sample is an exception to the general rule, for the Tapeats contains more iron oxides than the underlying weathered Archean. This is evident even in the photograph (Pl. 5, fig. 2) and is believed to be due to removal of the more deeply weathered detritus from the Archean and incorporation in the Tapeats of iron-rich material derived from a near-by area.

Not all of the lower Tapeats beds are composed of detritus derived from the underlying rocks, for in many places they consist largely of foreign material. On Tapeats Creek the Tapeats rests on Shinumo quartzite, but the basal beds consist predominantly of large angular quartz fragments derived from the Archean. Likewise, at mile 96 the basal Tapeats consists entirely of pegmatitic material, and the underlying rocks contain no pegmatite, although they are not far distant.

Tapeats grit, conglomerate, and breccia.—Lenses of grit and conglomerate or breccia are abundant in the lower Tapeats. The grits are usually arkosic and consist of angular quartz and feldspar fragments such as might have been derived from a pegmatite. The conglomerate lenses or pockets consist predominantly of a ferruginous sandy matrix containing angular quartz fragments half an inch to 8 inches

TABLE 2.—*Iron oxides in basal Tapeats and in underlying weathered material*

Locality	Material	Iron oxide (Per cent)
Garden Creek trail	Basal Tapeats	2.80
	Underlying Archean	3.30
Garden Creek Indian ruins	Basal Tapeats	4.90
	Underlying Archean	3.70
Ninetyone Mile Creek	Basal Tapeats	4.40
	Underlying Archean	12.80
Mile 117½	Basal Tapeats	3.20
	Underlying Archean	7.20

in diameter, probably derived from pegmatite dikes and quartz stringers or veins. Good examples of such deposits have been observed at the Garden Creek Indian ruins (Pl. 5, fig. 2), at mile 117½, just below 120 Mile Rapids, at mile 127, and elsewhere. The lenslike form of the deposits is probably due to accumulation in local, shallow depressions on the sea floor.

Deposits of coarse material on the flanks of Ep-Algonkian monadnocks are also noteworthy. Usually, at the base of the monadnocks is an accumulation of large angular blocks which may be Ep-Algonkian talus or, more likely, material derived by wave erosion as the sea encroached on the flank of the monadnock. Higher in the section is more coarse material clearly derived from higher parts of the monadnock at a somewhat higher stand of the sea. These beach shingles, if they may be called such, consist of large angular blocks close to the monadnock and pass gradationally outward into material consisting of smaller and more rounded fragments until the conglomeratic nature is entirely lost. Noble (1910, p. 501, 527; 1914, p. 62; 1922, p. 38), Walcott (1883, p. 439; 1895, p. 327), Ransome (1917, p. 159), Hinds (1935, p. 27), and others have noted this relation. It has been observed on the flanks of many Ep-Algonkian monadnocks, one example of which is sketched in Figure 7. Dake and Bridge (1932, p. 630-636) report similar conditions in the Ozarks.

An unusual breccia near the base of the Tapeats has been observed in Ninetyone Mile Canyon just east of Trinity Creek in the Bright Angel quadrangle. The basal Tapeats consists of 2 to 5 feet of alternating, thin, brown and white sandstone beds (Pl. 4, fig. 2). Above the basal zone of thin beds is a coarse breccia, 5 to 20 feet thick, consisting

chiefly of large angular fragments of Shinumo quartzite up to 10 feet in diameter (Pl. 4, fig. 2; Pl. 6, fig. 1). A few scattered angular fragments of the Archean complex have also been noted. The breccia is of local extent as it does not appear on the east wall of Ninetyone Mile

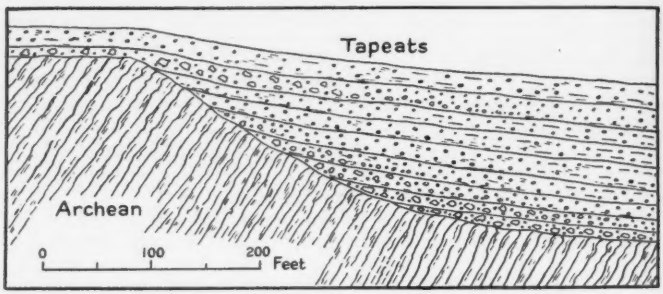

FIGURE 7.—*Tapeats breccia and conglomerate lenses on the flank of a monadnock*

Scale applies only to height of monadnock, not to fragments in Tapeats, Mile 138½.

Canyon nor on the walls of Trinity Creek (Campbell, personal communication). A notable feature is the deformation of the underlying beds which occurred at the time the breccia was formed. Figure 1 of Plate 6 shows very well the nature of this deformation.

The size and shape of the breccia fragments, their composition, the lack of bedding or sorting, the local extent of the breccia, and the associated deformation are all thought to indicate a slide as the most likely mode of origin. The sequence of events pictured is as follows. The encroaching Tapeats sea lapped over the relatively flat Ep-Algonkian surface and against the southwest flank of a monadnock of Shinumo quartzite (Pl. 6, fig. 2). The detritus mantling the surface was reworked to form the relatively fine basal Tapeats. Wave erosion oversteepened the slopes of the monadnock to such an extent that a great mass of material broke loose, slid down into the sea, and moved southwestward out over the fine beds, deforming them locally. Waves and currents smoothed off the top of the slide and rounded some of the boulders on its upper surface; continued deposition of finer material buried the mass completely. A more thorough description of this slide breccia has been presented elsewhere (Sharp, 1940).

The basal part of the Tapeats sandstone is extremely variable in composition and grain size, a reflection of its local origin. It consists of regolith reworked more or less *in situ*, of weathered material introduced

FIGURE 1. SLIDE BRECCIA AT BASE OF TAPEATS, NINETYONE MILE CREEK
Breccia bed is about 15 feet thick.

FIGURE 2. VIEW NORTHWEST ACROSS INNER GORGE TO HEAD OF NINETYONE MILE CREEK
S—Face of Shinumo quartzite monadnock from which slide breccia (B) at base of Tapeats was derived. Ar—Archean complex. (Photograph by courtesy of National Park Service.)

FIGURE 1. MONADNOCK OF SHINUMO QUARTZITE, SHINUMO CREEK
Dashed line—trace of Ep-Algonkian surface, solid line—top of Tapeats, dotted line—present outline of monadnock. Tapeats wedges out between slope of the monadnock and overlying Bright Angel.

FIGURE 2. TAPEATS-ARCHEAN CONTACT, DEER CREEK
Tapeats wedges out against the flank of the monadnock.

EP-ALGONKIAN MONADNOCKS

from near-by outcrops, or of fairly fresh material introduced from more deeply eroded areas. Lenses and shingles of coarse material are common, particularly at the base or on the flanks of monadnocks, and one particularly fine example of slide breccia has been described.

EP-ARCHEAN AND EP-ALGONKIAN CLIMATIC CONDITIONS

GENERAL STATEMENT

The available evidence does not permit a differentiation between climatic conditions existing toward the close of the Ep-Archean and Ep-Algonkian intervals. For this reason and for the sake of brevity the Ep-Archean and Ep-Algonkian climates are considered together on the assumption that they were essentially the same; it is realized, however, that more sensitive evidence might indicate notable differences. Relatively little is known of pre-Cambrian climates because of the lack of suitable plant and animal remains. Recourse to physical features appears to be the only method of attack, and by this method some progress has been made, as, for example, the recognition of pre-Cambrian glaciations.

Noble (1910, p. 524; 1914, p. 81) has suggested that evidence of only slight chemical weathering on the Ep-Archean surface, the arkosic nature of the weathered material, and the red color of the Hotauta conglomerate matrix all indicate an arid climate. Hinds (1935, p. 14, 18, 20-21, 50) states that the preponderance of mechanical disintegration, evidence of slight chemical weathering, and the fresh feldspar in the weathered material on the Ep-Archean surface suggest either dry microthermal or tundra conditions; he adds (Hinds, 1935, p. 51) that conditions were probably about the same during Ep-Algonkian time. Various lines of evidence are here cited which lead to the conclusion that the climate during the closing phases of the Ep-Archean and Ep-Algonkian intervals was relatively humid.

EVIDENCE ON CLIMATIC CONDITIONS

Red color of weathered material.—Raymond (1927) has shown that red regoliths are more likely the result of warm, humid climates than of arid or semiarid conditions, and this appears to be a time-honored and commonly accepted view (Russell, 1889, p. 27, 46; Crosby, 1891; Barrell, 1908, p. 285, 287, 291; Glinka, 1927, p. 35-42; Hager, 1928, p. 920-921, 923; Twenhofel, 1932, p. 132, 278). Hematite is usually considered to be the major source of red color in such accumulations, but Posnjak and Merwin (1919) have demonstrated that turgite, which appears to be a solid solution of goethite and hematite with adsorbed and capillary water, is red. Furthermore, there are artificial amorphous

ferric oxide compounds containing dissolved and adsorbed water which may have as yet unrecognized counterparts in nature. Therefore, it is possible that red color does not necessarily indicate complete dehydration as has been assumed in some cases. Red regoliths such as those on the Ep-Archean and Ep-Algonkian surfaces are formed under moderately warm, humid climates.

Feldspars in weathered material.—Most of the feldspar in the weathered material is orthoclase or microcline; plagioclase, being much less resistant to weathering, has been destroyed. Furthermore, in the most maturely weathered material either on the surface or in the overlying beds, the potash feldspars have also been eliminated by weathering. For this reason the arkosic nature of the deposits is attributed not to arid climate but to marine scouring and reworking, the assumption being that the maturely weathered material has been largely removed and locally incorporated in the overlying beds, and this is supported by field observation (Pl. 5, fig. 2). Only a few feet of weathered material would have to be removed to expose relatively unweathered potash feldspar, for as noted by Watson (1902, p. 311) residual weathered detritus on granite in Georgia, an area of deep chemical decay, contains relatively fresh potash feldspar at a depth of only 5 feet. Furthermore, not all the feldspar in the basal Tapeats need have come from the immediately underlying detritus, for some of it may have been imported from a near-by area from which the deeply weathered material had already been removed. This matter has been ably discussed by Pumpelly (1891, p. 210-214).

The question of the significance of clastic feldspar in sedimentary deposits as a climatic indicator appears to need reconsideration. Barton's (1916) study is justly recognized as classical, but such statements as those of Reed (1928) concerning the abundance of feldspar in deposits formed under humid conditions indicate that the entire story is not known. Under any circumstances, the arkosic nature of the material on the Ep-Archean and Ep-Algonkian surfaces and in the overlying beds does not necessarily indicate an arid climate.

Insoluble residues.—Hilgard (1912, p. 384) reports that the most striking difference between soils of arid and humid regions is the relative percentages of insoluble residue. The insoluble residue in the average of 696 analyses of humid region soils is 84 per cent, and it is 69 per cent in an average of 573 analyses of soils from arid regions (Hilgard, 1912, p. 377). The average insoluble content of nine different samples of maturely weathered material from the Ep-Archean and Ep-Algonkian surfaces, determined by the Hilgard method (Wiley, 1906,

p. 392) is 88 per cent. Only the most maturely weathered material was used in order to eliminate errors caused by incomplete weathering, and the entire procedure is open to criticism to the extent that the materials analysed may not be true zonal soils. The results are interpreted as indicating relatively humid climate.

Iron oxide content of weathered detritus.—One of the principal results of weathering has been the formation of iron oxides and hydrates and in some cases a residual enrichment in these compounds. Clarke (1924, p. 490-491) gives analyses of nine fresh and weathered igneous and metamorphic rocks from humid climates, chiefly eastern United States. In most of these analyses the iron, calculated as oxides, remains essentially constant or increases during weathering. No similar data from arid regions have been found for comparison, but the close similarity between the weathered material from the Grand Canyon and that described by Clarke is at least suggestive of humid conditions during the weathering of the Grand Canyon material.

Type of weathering.—The weathering processes have already been considered in detail, and the dominance of chemical over mechanical weathering has been noted. Usually, chemical weathering is assumed to be dominant in humid climates and mechanical weathering in arid climates. Blackwelder (1927, p. 127-128) believes that chemical weathering is dominant under arid as well as humid conditions. However, it is questionable whether the types and degree of chemical weathering described in this article would occur under arid conditions, and certainly mechanical weathering has not dominated as stated heretofore (Noble, 1910, p. 524; 1914, p. 81; Hinds, 1935, p. 14, 20-21).

Caliche.—The significance of caliche or lime-rich zones associated with erosion surfaces has been discussed by Woolnough (1928), Wolfanger (1929), Price (1933), and Bryan and McCann (1938, p. 5-6). The general conclusion reached is that caliche is produced by extended weathering upon any type of rock under arid or semiarid conditions. Price (1933, p. 500-501) distinguishes various types of caliche of which one—the "old caliche"—is particularly characteristic of aridity. Woolnough's (1928, p. 893) "duricrust" is a particular type of caliche which is developed on a peneplain under climatic conditions marked by sharply defined alternations of saturation and desiccation.

The absence of caliche on the erosion surfaces is in itself not absolute proof that they were formed under humid conditions, for such a caliche might have been removed by marine erosion or perhaps leached by modern weathering. However, Price's (1933, p. 501) "old caliche" or Wool-

nough's (1928, p. 893) "duricrust", both supposedly indicative of arid or semiarid conditions, would probably be tough enough to withstand marine erosion and modern leaching at least locally. The absence of any traces of such accumulations is suggestive that conditions favorable to their formation did not exist during the Ep-Archean and Ep-Algonkian intervals.

Miscellaneous features.—The penultimate and ultimate forms of the arid erosion cycle consist essentially of coalescing pediments with or without intervening areas of basin fill depending upon the nature of the drainage. Surfaces of pedimentation even in advanced stage are not so smooth and flat as the Ep-Archean surface or parts of the Ep-Algonkian surface. Furthermore, limestone, usually moderately resistant to weathering under arid conditions, was less resistant than other rocks. The resistant rock types on the Ep-Algonkian surface were those which are usually resistant under normal humid or semihumid conditions.

Barrell (1917, p. 760) and Woolnough (1928, p. 888) have suggested that the water table beneath peneplains should be close to the surface. That it is at a correspondingly shallow depth in arid regions does not seem reasonable. Pre-Cambrian weathering of any notable degree extends to an average depth of 10 to 15 feet, and this is relatively shallow compared to depths of weathering cited by Merrill (1897, p. 276-277) and Twenhofel (1932, p. 17) from areas of somewhat greater relief in humid climates. The amount of weathered material removed by the encroaching seas is not definitely known but appears to be moderately small judging from the amount of weathered material in the overlying sedimentary beds. The lack of evidence of an exceptionally thick weathered layer seems to indicate that the water table must have been relatively near the surface, a relation likely under at least moderately humid conditions.

A common feature of arid regions is the thick deposits of alluvium filling undrained basins. Remnants of such fills on the Grand Canyon surfaces might be evidence of at least moderate aridity. Their absence is not proof of a humid climate but it at least suggests that both surfaces were drained by through-flowing streams. It is possible that wind may have removed alluvial material, but no evidence of wind action was found on either surface.

The probable lack of vegetation on these surfaces is undoubtedly an important factor but one which cannot be easily evaluated. The lack of vegetation might account for certain types of weathering, for the lack of thick normal soils, and would favor wind action.

Thus, the available evidence indicates that in the Grand Canyon area

the climate during at least the latter parts of the Ep-Archean and Ep-Algonkian intervals was not arid as heretofore suggested but was probably moderately warm and humid.

ORIGIN OF SURFACES

PROCESSES AND CONDITIONS OF PENEPLANATION

The concept of a peneplain as the penultimate product of subaerial erosion is now accepted after an initial period of skepticism and doubt (Davis, 1889; 1896; 1899a; 1899b; Tarr, 1898). The Grand Canyon surfaces, although buried beneath marine deposits, are believed to be of subaerial origin, chiefly because of the mantle of weathered debris upon them. It could be argued, perhaps, that the surfaces had been formed by marine planation and that they had been exposed to weathering prior to encroachment of the Algonkian and Tapeats seas. The topographic relief of at least the Ep-Algonkian surface weakens this theory, and the wide extent of these surfaces in the Grand Canyon and elsewhere in Arizona favors a subaerial origin, for as noted by Wentworth (1927, p. 131) and Meyerhoff and Hubble (1928, p. 341-344) marine platforms cut on hard rocks, such as those of the Archean complex, are likely to have a limited areal extent. Even the surfaces of marine planation mentioned by Johnson (1919, p. 228-231) are not so extensive as these surfaces.

Apparently no one has questioned seriously the subaerial origin of the Grand Canyon surfaces, although Davis (1900, p. 253; 1901, p. 176; 1909b, p. 350) has suggested that the Ep-Archean surface may be due at least partly to marine planation, and Moore (1927, p. 273) has noted that the smoothness of these surfaces might be attributed to marine planation, perhaps supplementing subaerial peneplanation.

The extreme smoothness of the Ep-Archean surface appears to be due to the essentially homogeneous nature of the underlying rocks and the long time available for erosion. The rocks beneath the Ep-Archean surface, although heterogeneous in local outcrop, as a group offer far more uniform resistance than do the Algonkian rocks exposed on the Ep-Algonkian surface. Erosion by the encroaching Algonkian sea appears to have been relatively minor as far as adding to the general smoothness of the surface, for in most places not even all the weathered rock has been removed by marine erosion.

It appears that removal of waste by running water, although decreasing in importance toward the close of the erosion cycle, has continued in some degree practically up to the end. Mass movement of material, particularly slow creep, has probably increased as a regolith was developed and has removed many of the remaining topographic ir-

regularities. Evidence of surface creep has been noted on the Ep-Algonkian surface. A thicker regolith over low spots on the surfaces protected those areas, while the higher places not protected by such a cover were lowered by various erosive agents. It is possible that solifluction and other types of frost action were effective, although no evidence has been found to support this view. Wind may have had a small part in the removal of fine material from this area, although no evidence of wind action was found.

The relative efficacy of mechanical and chemical weathering probably changed during the cycle. The effects of chemical weathering cannot be overlooked in an evaluation of the processes active on the surfaces in the final stages. The probable lack of vegetative cover removes a factor which is highly significant in the formation of most present-day erosion surfaces.

The Grand Canyon surfaces are truly the product of subaerial erosion, and weathering, mass movement, and running water are believed to be the major processes active in their formation. The Ep-Archean surface is a particularly fine example of a surface of extremely low relief produced by subaerial processes.

THICKNESS OF ROCK REMOVED AND DURATION OF EROSION INTERVALS

The thickness of rock removed from the Ep-Archean surface is not known, and no reliable method of determining it appears applicable. The coarse texture of the Archean igneous intrusives shows that they crystallized at some depth, but whether that depth was 2000, 5000, or 15,000 feet cannot be determined. Noble (1934, p. 27) states that mountains 20,000 feet high were worn down to form the Ep-Archean surface, but his evidence for this is not given. Lawson (1934) has used the principle of isostatic rise through lightening of the rock column by erosion to calculate the length of time and the thickness of rock which would have to be removed before final isostatic equilibrium would be attained. By this method he has calculated that 21 million years were required to remove 12.24 kilometers (40,200 feet) of rock to produce the Ep-Archean surface of the Canadian Shield. The same calculation and figures might be applied to the Ep-Archean surface in the Grand Canyon, but the method is founded to such an extent on arbitrary assumptions that the final results remain chiefly a subject of speculative interest as indicated by Lawson himself (1934, p. 1072) and Hinds (1935, p. 4).

The minimum thickness of Algonkian beds removed by erosion during the Ep-Algonkian interval can be calculated easily for the Shinumo quadrangle from the data given on Noble's map and sections (1914, Pl. 1). Using a thickness of 12,000 feet for the Algonkian beds, and an average

dip of 20 degrees, the thickness of Algonkian rocks which was removed amounts to 12,800 feet. Hinds (1935, p. 44) reports that approximately 13,000 feet of Algonkian rocks have been removed by Ep-Algonkian erosion in the area studied by his parties. In many places a considerable thickness of Archean rock below the Algonkian beds has also been removed. A minimum figure for the thickness of Archean rock eroded can be obtained by projecting the inclined base of the Algonkian beds above the Ep-Algonkian surface until it is truncated by a fault or passes under the Paleozoic cover. Using Noble's section CC′ (1914, Pl. 1) as a basis, the thickness of Archean rocks removed by Ep-Algonkian erosion has been calculated as approximately 2000 feet. Therefore, the total thickness of rock eroded during the Ep-Algonkian interval is close to 15,000 feet. This is a conservative calculation, and in many places the thicknesses were probably considerably greater.

Ashley (1935, p. 1403) estimated that the high ridges of the Appalachians have been lowered approximately 100 feet for each million years that they have been exposed. Lawson (1934, p. 1071) combined the studies of Dole and Stabler (1909) and of Humphreys and Abbot (1861) to arrive at a figure of lowering of the Mississippi River hydrographic area of one foot in 5415 years or roughly 185 feet per each million years. If 150 feet per million years is used as an approximate mean figure, removal of 15,000 feet of rock to form the Ep-Algonkian surface would require 100 million years. Obviously this figure is only an approximation, for variations in climate, topographic relief, rock types, and stage of the erosion cycle are all unevaluated. Furthermore, if the surface were barren of vegetation, the rate of erosion may have been considerably greater and the elapsed time correspondingly shorter.

By no means should the inference be drawn that 100 million years of erosion are required to produce a peneplain, for it is well known that many peneplains have been formed in a small fraction of that time. There is no way of knowing how many complete or incomplete erosion cycles are represented by the erosion interval between the initial phases of the Grand Canyon disturbance and the deposition of the Cambrian Tapeats sandstone. It is realized that this calculation must be regarded with some skepticism because of the assumptions upon which it is based. However, such calculations are not without academic interest at least.

EFFECTS OF MARINE EROSION ON SURFACES

GENERAL STATEMENT

The rate of encroachment of a shore line and the associated shallow-water zone over a land surface is determined by the rate of downwarping or eustatic rise of the sea and the relief of the submerged surface. Ordi-

narily wave action is effective to an average depth of 50 to 60 feet and in times of exceptional storm it may extend as deep as 200 to 600 feet (Johnson, 1919, p. 72-80). Therefore, it is difficult to imagine submergence of any land surface as being rapid enough to prevent at least some reworking of the weathered mantle upon that surface. If the rate of submergence were slow and the sea shallow, marine erosion should modify the land surface to a greater degree than if the submergence were rapid and the sea deep, a matter discussed briefly by Twenhofel (1932, p. 144-145) and Longwell (1925, p. 104-106). The availability of sediments with which the sea floor could be partly protected from erosion cannot be easily evaluated but must be significant in some cases (McKee, personal communication). Modification of the Grand Canyon surfaces by marine processes has been so slight that submergence of these surfaces must have been moderately rapid. The modifications have been of two types: (1) reworking and transportation of weathered detritus and the topographic changes resulting therefrom, and (2) erosion of monadnock slopes.

MODIFICATION OF EP-ARCHEAN SURFACE

The Ep-Archean surface shows only minor modification by marine processes. This is probably due partly to its low relief which permitted rapid submergence with only a slight relative movement of land and sea and partly to the quiet sea indicated by the Bass limestone.

Only part of the residual Archean regolith has been reworked, and in only a few places has marine erosion entirely removed all the weathered rock. The original thickness of the weathered layer is not known, but the thinness of the reworked weathered material suggests that it was not great. Thus, even though the Algonkian sea has removed part of the weathered mantle, this removal has not greatly affected the topographic nature of the Ep-Archean surface, and marine erosion is responsible only in minor part for the extreme smoothness of the Ep-Archean surface.

MODIFICATION OF EP-ALGONKIAN SURFACE

The effects of marine erosion on the Ep-Algonkian surface are more impressive. Nearly everywhere some of the Ep-Algonkian regolith has been reworked, and in places all the weathered mantle has been removed, and relatively fresh rock has been eroded. Presumably, the higher parts of the Ep-Algonkian surface have been subjected to the most intense and prolonged erosion and, therefore, have been somewhat lowered with a resultant smoothing of the surface. Furthermore, as suggested by McKee (Wheeler and Kerr, 1936, p. 4-5) subaerial erosion has probably reduced the higher monadnocks in the interval between initial encroachment of the Cambrian sea and complete submergence in late Tapeats or Bright Angel time.

The most impressive evidence of marine erosion is seen on the flanks of Ep-Algonkian monadnocks. As noted by Noble (1910, p. 527; 1914, p. 62) wave erosion on these monadnocks has sharpened their slopes and considerably modified their outline. Figure 8 is a sketch of a buried

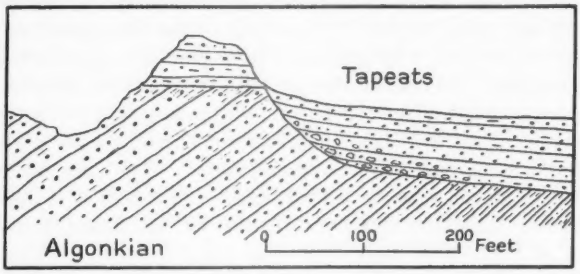

FIGURE 8.—*Buried monadnock modified by marine erosion, Hotauta Creek*
Sharp angle between steep slope and flat top partly caused by marine erosion.

monadnock on the south wall of Hotauta Canyon. Noble (1914, p. 62) described the steep face of this monadnock as a sea cliff. Perhaps more significant than the steep slope is the flat top of the monadnock and the angle between the two. The outline of this monadnock has been modified by marine erosion both by steeping of the slope and by smoothing of the top upon further submergence.

The coarse material derived by wave erosion on the monadnocks has been worn, sorted, and deposited in lenses in which the fragments become finer and more rounded and the whole lens thinner at increasing distances from the monadnock. Many excellent examples of successions of such lenses or beach shingles, presumably formed by progressive encroachment of the sea against the flanks of the monadnocks, have been observed, and one is sketched in Figure 7. At mile 138 small terracelike platforms were noted on the slopes of a monadnock. If these features are truly terraces, they may indicate short pauses in the submergence of the Ep-Algonkian surface.

One of the most spectacular results of marine erosion is the slide breccia at the base of the Tapeats in Ninetyone Mile Canyon (Pl. 6, fig. 1). This breccia has been derived from an oversteepened sea cliff carved by wave erosion on a monadnock of Shinumo quartzite. Coarse breccias were observed at the base of practically all monadnocks of any size, but this was the only example of a slide breccia noted. The initial Tapeats sea must have been a shallow sea above whose surface rose numerous islands. Some of these islands were so high that they were not completely submerged until well into Bright Angel time.

SUMMARY

The buried Ep-Archean and Ep-Algonkian erosion surfaces of the Grand Canyon are excellent examples of well-preserved peneplains. The maximum relief of the Ep-Archean surface is less than 50 feet, and this extreme smoothness is probably due to long erosion and to uniformity of the underlying rocks. At least 95 per cent of the Ep-Algonkian surface has a relief of less than 150 feet, and a large part is essentially flat. The remainder consists of monadnocks, underlain by massive Archean granite or Shinumo quartzite, which rise to a maximum height of 800 feet.

Both surfaces show evidence of subaerial weathering, and this weathered mantle has been studied in considerable detail. The greatest depth of weathering is 10 feet beneath the Ep-Archean surface; it is 50 feet beneath the Ep-Algonkian surface, but only the uppermost 5 to 10 feet is greatly altered. The following progressive changes with increasing intensity of weathering have been noted. Biotite changes first to a green chlorite and then to a colorless chloritelike material in which are dispersed small grains of iron oxide. Eventually, biotite is entirely eliminated, leaving hydrated iron oxide. Feldspars show some kaolinization even at considerable depth, and plagioclase is almost entirely decomposed in an early stage of the weathering. Potash feldspar has been more resistant, although in some places it too has been entirely decomposed. Quartz and muscovite have been relatively unaffected and show a residual enrichment. Quartz and muscovite with hydrated iron oxides and clay constitute most of the residual regolith produced by extended weathering; the amount of iron oxides ranges from 0.9 to 12.8 per cent. In a few places a residual concentration of iron oxide in the weathered zone appears certain, and it is suspected in other localities. Insoluble residues constitute an average of 88 per cent of the most maturely weathered material. The evidence of chemical weathering on these surfaces is far greater than the evidence of mechanical weathering.

Several lines of evidence offer proof that the weathering on the Grand Canyon surfaces is ancient. Perhaps the most convincing is the incorporation of the weathered detritus in the basal part of the overlying sedimentary beds.

The red color of the weathered detritus, the evidence of extensive chemical weathering, the high percentage of insoluble residues, the lack of soil caliche, and other types of evidence are believed to indicate a relatively humid climate during the closing phases of the Ep-Archean and Ep-Algonkian erosion intervals. The probable absence of plant life is admittedly a factor which cannot be easily evaluated.

The Grand Canyon surfaces, although buried beneath marine deposits, are believed to be the product of subaerial erosion and not marine plana-

tion, chiefly because of the evidence of subaerial weathering on both surfaces. No evidence of wind action was noted, and it seems likely that weathering, mass movements, and running water were the major factors producing the low relief of both surfaces.

Calculations are made showing that in some places at least 15,000 feet of rock has been eroded from the Ep-Algonkian surface. Using a composite figure for the rate of lowering of land by erosion, it is calculated that roughly 100 million years were required to produce the Ep-Algonkian surface.

It is concluded that marine processes have modified the topography of the surfaces only to a minor degree. The major effects of marine erosion have been (1) a reworking of part of the weathered material and its incorporation in the overlying beds, (2) a slight levelling of the surfaces, and (3) a more noticeable and spectacular steepening of the slopes of monadnocks on the Ep-Algonkian surface. The face of at least one such monadnock has been so oversteepened by marine erosion that it has broken away in the form of a slide.

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